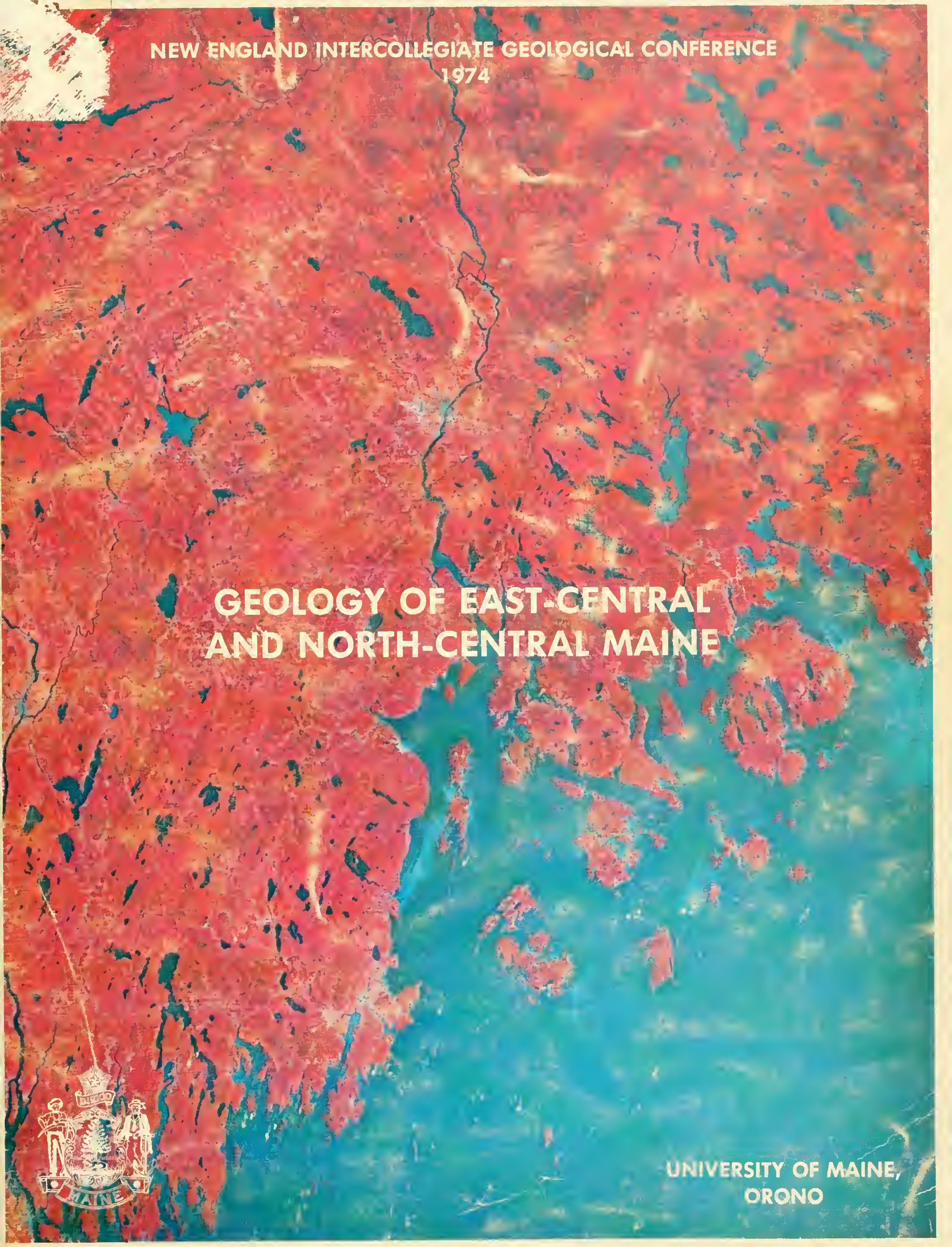


NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

1974

GEOLOGY OF EAST-CENTRAL AND NORTH-CENTRAL MAINE

UNIVERSITY OF MAINE, ORONO

NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

66TH ANNUAL MEETING

GUIDEBOOK FOR FIELD TRIPS IN EAST-CENTRAL AND NORTH-CENTRAL MAINE

October 12 and 13, 1974

Philip H. Osberg, Editor

Orono, Maine

FOREWORD

The 1974 New England Intercollegiate Geological Field Conference is headquartered at Orono for the purpose of examining the geology of north-central and east-central Maine. Excursions have been arranged to cover the stratigraphy, sedimentology, structural geology, metamorphic and igneous petrology, mineral deposits, and glacial geology of the region.

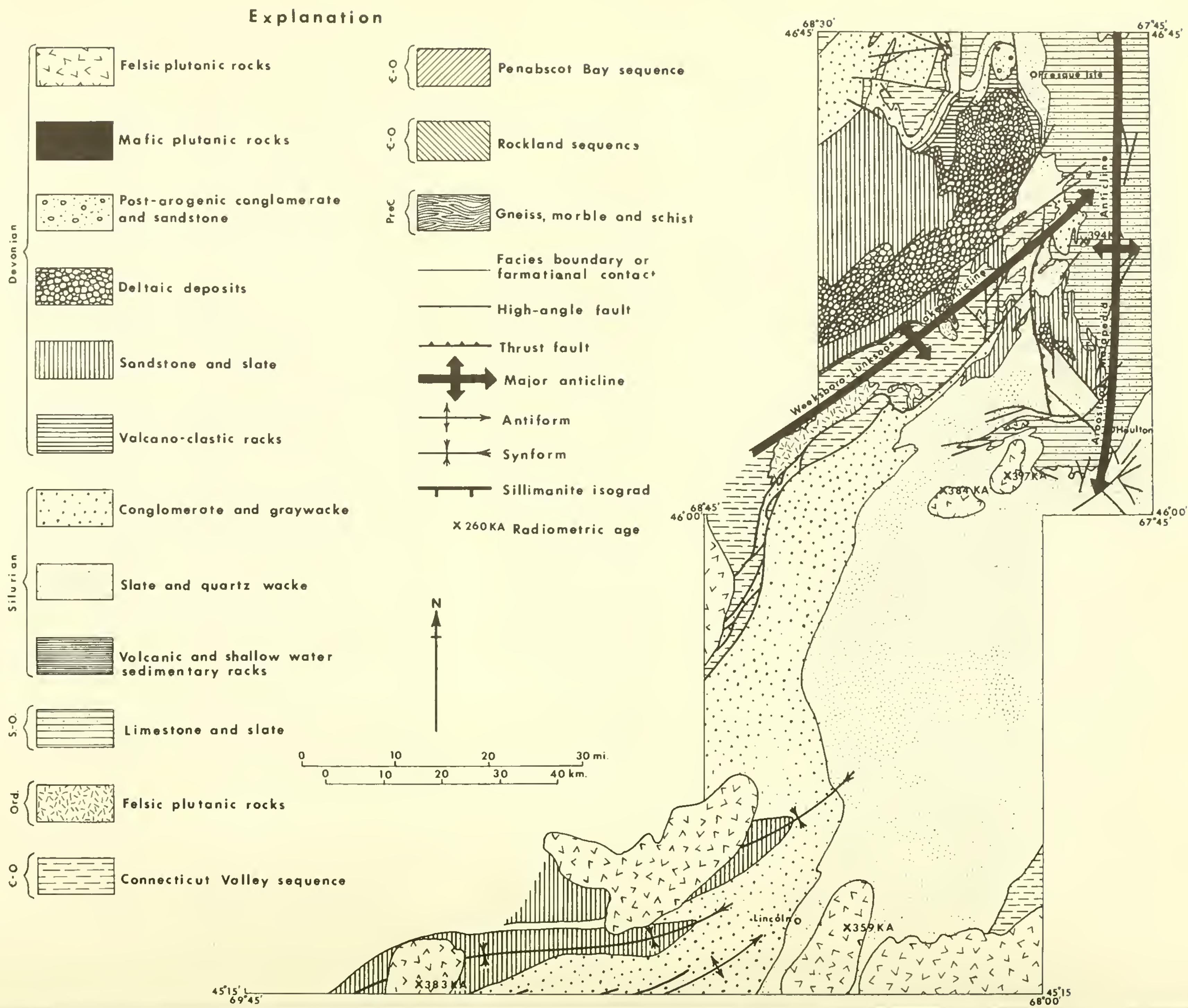
In order to provide participants with an appropriate background to better appreciate the field trips a broad overview of the regional geology is presented. To this end a tectonic map showing the pertinent geologic features appears as Figure 1.

In the construction of Figure 1 and in the description that follows, I have drawn heavily on the work of my colleagues at this meeting and on that of many others who have worked in Maine and elsewhere in the northern Appalachians. These people have taken part in stimulating and sometimes heated discussions on the geology of north-central and east-central Maine and have had a major part in developing my understanding of the region. No attempt has been made to credit these individuals in the text, although the author can claim few of the ideas presented as being originally his. Instead, a bibliography is listed below which represents the primary sources for many of the data presented in the Foreword.

Precambrain rocks are confined to the Liberty-Orrington anticline and a small horst in Penobscot Bay (Fig. 1). The Precambrian rocks of Penobscot Bay consist of metamorphosed schist and marble intruded by pegmatites for which radiometric ages indicate a date of approximately 600 my (Stewart, this volume). Feldspathic gneisses and schists in the core of the Liberty-Orrington anticline have been designated as Precambrian because they have been metamorphosed to high grade prior to the intrusion by granites of Ordovician age (Wones, this volume) and to the development of faults and later metamorphism.

Cambro-Ordovician rocks have been divided into three sequences: the Connecticut Valley sequence consisting of noncarbonaceous schist and metasandstone, metavolcanic rocks, black pyritiferous schist, and metagraywacke; the Penobscot Bay sequence consisting of metaconglomerate, noncarbonaceous schist, and black pyritiferous schist (Metavolcanics make up only a minor part of this sequence.); and the Rockland sequence consisting of noncarbonaceous schist, quartzite, marble, and metaconglomerate. The Penobscot Bay and Rockland sequences are juxtaposed along faults, but these two sequences are nowhere contiguous with the Connecticut Valley sequence. Exposures of the Connecticut Valley sequence lie northwest of a line passing through Lewiston and Bangor (Fig. 1), whereas the Penobscot Bay and Rockland sequences are restricted to exposures southeast of that line.

The Connecticut Valley sequence ranges in age from Cambrian to Middle Ordovician (Caradoc). The Rockland sequence, although different in lithic sequence, contains fossils that have been assigned to the Caradoc as well.


Explanation
Devonian
Felsic plutonic rocks
Mafic plutonic rocks
Post-arogenic conglomerate and sandstone
Deltaic deposits
Sandstone and slate
Valcano-clastic rocks
Silurian
Conglomerate and graywacke
Slate and quartz wacke
Volcanic and shallow water sedimentary racks
S-O.
Limestone and slate
Ord.
Felsic plutonic rocks
€-O
Connecticut Valley sequence
€-O
Penabscot Bay sequence
€-O
Rockland sequencs
PreC
Gneiss, morble and schist
Facies boundary or farmatianal contact
High-angle fault
Thrust fault
Major anticline
Antiform
Synform
Sillimanite isograd
X 260 KA Radiometric age
N
0 10 20 30 mi.
0 10 20 30 40 km.
68°30'
46°45'
67°45'
46°45'
Presque Isle
Anticline
394KA
Weeksboro-Lunksoos Lake Anticline
Aroostook-Matapedia
Houlton
X397KA
X384KA
68°45'
46°00'
46°00'
67°45'
Lincoln
X359KA
X383KA
45°15'
69°45'
45°15'
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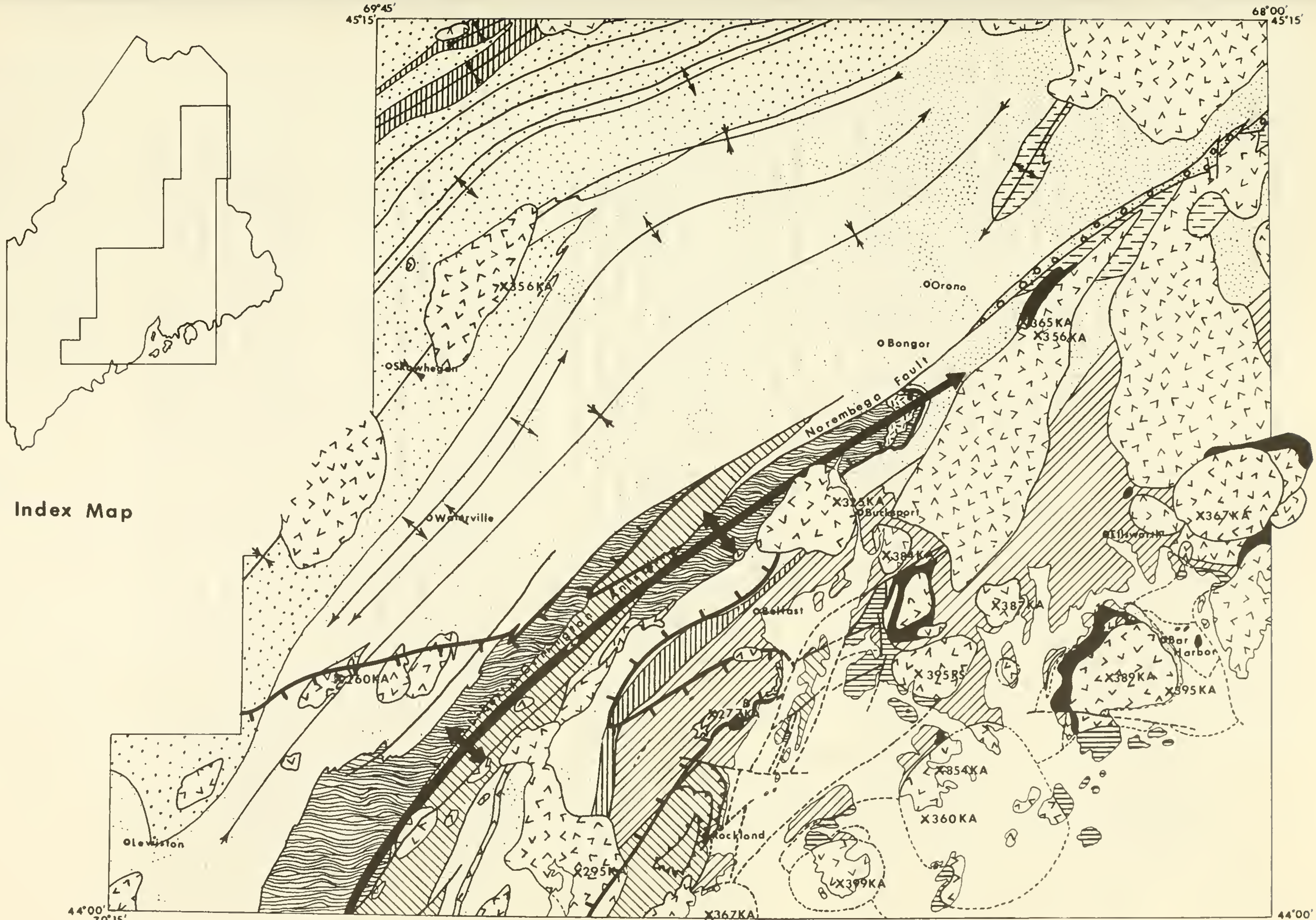

Figure 1. Tectonic map of east-central and north-central Maine.

Presumably the Rockland sequence ranges in age from Cambrian to at least Caradoc. The Penobscot Bay sequence contains no fossils, but its uppermost unit has been traced in reconnaissance into the Oak Bay Formation of Arenig age near St. Stephen, N.B.

The presence in close proximity of the three different sequences provides a basis for interesting speculations. The three sequences could have been deposited in the same basin as contiguous sediments with intertonguing relationships. The present juxtaposition of the Penobscot Bay and the Rockland sequences would be explained by a combination of thrust and high-angle faulting. Alternatively, it is tempting to correlate the Penobscot Bay sequence on the basis of stratigraphy to the Cambro-Ordovician sections at St. Johns, N.B. and to those on the Avalon Peniusular in Newfoundland. On this basis the Penobscot Bay sequence may be Euro-Afrian and the Connecticut Valley sequence may be typical of American deposition during Cambro-Ordovician time. According to this speculation the two sequences were brought together by plate motions in post-Middle Ordovician time. In this case the trace of the suture separating the two continents must lie approximately along the line between Lewiston and Bangor (Fig. 1). The Rockland sequence could represent a facies variant of either the Connecticut Valley or the Penobscot Bay sequence which has been faulted into its present position.

Silurian flysch-like sedimentary rocks dominate the geology of north- and east-central Maine. These deposits are in contact with different lithic units of the pre-Silurian section at different places. Some of these contacts have been interpreted as normal contacts, others as faults, and many have relationships that are incompletely understood. However, the regional relationships suggest the possibility of an unconformity at the base of the Silurian section, particularly in northwest and southeast directions toward the margins of the Silurian depositional basin. It is to be noted that if the Silurian section (deposited in a narrow trough) is unconformable on pre-Silurian rocks of both the American and the Euro-African plates, the continental segments of the two plates must have made contact before Silurian time.

The Silurian deposits have been divided into three facies: a conglomerate-graywacke facies, a slate-quartz wacke facies, and a limestone-slate facies. These rocks exhibit graded bedding, cut-and-fill features, flutes, and slump structures. The arrangement of the facies (Fig. 1) along with observations of a limited number of current direction indicators suggest that the major source for Silurian sediments was from the northwest. The limestone-slate facies in the vicinity of Presque Isle and Houlton partly interfingers and partly underlies the slate-quartz wacke facies. The lower part of the limestone-slate facies is Ordovician.

To the northwest of the basin containing flysch-like sedimentary rocks, the Silurian is thin and consists of shallow water conglomerates, limestones, limy slates, and felsic to mafic volcanic rocks. Rocks belonging to this shallow water facies crop out along the northwest flank of the Weeksboro-Lunksoos Lake anticline (Fig. 1). Somewhat similar rocks are exposed in the outer islands of Penobscot Bay. These rocks are likewise of shallow water origin consisting of conglomerate, sandstone,

shale, limestone, and, east of the map in Figure 1, of felsic to mafic volcanics.

The distribution of facies in rocks of Silurian age indicates that the basin containing flysch-like deposits which trends northeast through Maine (Fig. 1) is bounded both northwest and southeast by thin, shallow water, clastic and volcanic sequences. These thin marginal basin deposits are thought to have been formed on geanticlines underlain by Lower Paleozoic and Precambrian rocks. The northwestern margin may have been bounded by thrust or high-angle faults that were continuously active in the Silurian to provide a source area for the conglomerate-graywacke deposits which border it to the southeast.

Stratified rocks of Early Devonian age (primarily Siegenian) are predominantly flysch-like in character, although a volcano-clastic and a post-orogenic conglomeratic facies are also recognized. The flysch-like rocks include (1) the cyclically bedded sandstone-slate sequence that comprises the great expanse of Seboomook Formation of north-central Maine and (2) the fragmented parts of deltaic deposits consisting mostly of graywacke and sparse conglomerate. The cylically bedded sandstone-slate sequence partly underlies and partly interfingers toward the east with the deltaic deposits (Fig. 1). These relationships combined with some current direction data suggest that the Lower Devonian flysch-like sediments were derived from the east with the deltaic deposits having a regressive relationship to the cyclically bedded sedimentary rocks.

Early Devonian (Geddinian?) volcano-clastic rocks occupy much ground in the vicinity of Penobscot Bay and the outer islands to the east (Fig. 1). This sequence includes felsic and mafic flows, agglomerates, and sparse sandy units with a thickness far less than that of the flysch-like sequence. Volcano-clastic rocks of identical age crop out near Presque Isle where they underlie and interfinger with what may be equivalents of the lower parts of the flysch-like facies.

Post-orogenic conglomerates of Middle Devonian age unconformably overlie older rocks south of Presque Isle and east of Orono along the Normbega fault (Fig. 1).

The distribution of sedimentary rocks within the flysch-like facies indicates that the basin of deposition for Lower Devonian time lies within and to the northwest of the area of Figure 1. The deltaic deposits have an off-lap relationship to the cyclically bedded sandstone-slate deposits and indicate a geanticlinal source to the east and southeast. A volcano-clastic sequence developed on the geanticline contemporaneously with the deposition of flysch-like sediments to the northwest. It is not clear whether the volcanic rocks at Presque Isle were once contiguous with those of Penobscot Bay along a north-trending geanticline or whether they were deposited along a northeast-trending volcanic ridge within the basin of flysch-like deposition.

Structural features of regional significance include those folds and faults shown in Figure 1. Several major anticlines are shown along with the traces of smaller folds in areas where they have been delineated.

The oldest folds recognized are east-facing recumbent folds with amplitudes measured in thousands of feet. Such folds have been mapped only in the coastal region, but observations in central Maine suggest that recumbent folds are also present there. No recumbent structures have been reported from north-central Maine.

Upright, isoclinal folds are prevalent throughout the region and overprint the recumbent folds to an extent that makes observation of the recumbent folds difficult. Mesoscopic isoclinal folds have amplitudes of tens of feet, but these mesoscopic folds must be in the limbs of larger folds with wavelengths measured in miles and amplitude of the order of 1000-2000 feet in order to explain the mapped distribution of lithic units. Forces that first compressed and flattened the isoclinal folds then buckled the section thickened by isoclinal folding producing the long, open folds. The axial traces of these folds mostly trend northeast. The Liberty-Orrington anticline and the Weeksboro-Lunksoos Lake anticline in large part were formed in this folding.

More-or-less upright folds with north-trending axial surfaces overprint all older folds. This folding is manifest only as minor folds in central and coastal Maine. No major folds with styles and orientations appropriate to this deformation have been mapped in this area, however, the Aroostook-Matapedia anticline in north-central Maine (Fig. 1) has a trend that is appropriate for it. Moreover, its trend is at a large angle to that of the Weeksboro-Lunksoos Lake anticline which is interpreted as an earlier fold.

The traces of the major folds define sigmoidal patterns in east-central Maine at the lattitude of Skowhegan and in north-central Maine west of Houlton (See Pavlides, this volume). These patterns may be due to late flexure of the axial surfaces of earlier folds; may be due to changes in the stress field with geography, perhaps related to the configuration of the basement; or it may be due to interference patterns of northeast-trending and north-trending folds.

The three groups of folds with distinctly different styles and orientations deform the Silurian rocks of central Maine and rocks that have tentatively been assigned to the Devonian in coastal Maine. The recumbent folds and the northeast-trending folds are cut by plutons that have been dated as Middle Devonian. On this basis the recumbent folds and northeast-trending folds formed in Lower Devonian time. Relationships between the north-trending folds and the plutons is not clear, and it is possible that they post-date the intrusion of the plutons. The pre-Silurian rocks probably contain structures that predate those described above, but the heavy overprint of Devonian deformation has made their recognition difficult.

In addition to folds, thrust faults and many high-angle faults have been delineated on Figure 1. Two thrust faults have been mapped in the vicinity of Rockland. The earlier thrust fault is related to recumbent folding, whereas the younger thrust fault is related to more upright folding. The thrust fault indicated northwest of Houlton may have been formed as a result of the interaction of the Aroostook-Matapedia and Weeksboro-Lunksoos Lake anticlines. The thrust faults are thought to be the same age as the folds with which they are associated.

High-angle faults are numerous both in north-central and in coastal Maine (Fig. 1). Such faults may be more abundant over much of the region, but poor exposure has made their delineation difficult. The high-angle faults in north-central Maine are mostly due to adjustments resulting from the interaction of the north-trending Aroostook-Matepedia anticline across the northeast-trending Weeksboro-Lunksoos Lake anticline. The high-angle faults mapped in the coastal region separate a series of horsts and grabens, although some of the boundary faults may have considerable strike-slip movement (Stewart, this volume). The high-angle faults have at least three ages: the oldest were formed prior to the intrusion of Middle Devonian plutons; those of intermediate age cut the plutons but do not offset isograds of the Permian (?) regional metamorphism; the youngest faults post-date the Permian (?) regional metamorphism and the intrusion of Triassic (?) mafic dikes.

Plutonic rocks of various compositions and ages occur in north-central and east-central Maine (Fig. 1). The older plutons intrude the Cambro-Ordovician sequences but not the Siluro-Devonian section, and on the basis of a few radiometric dates are considered to be of Ordovician age (Wones, this volume). A large Ordovician pluton of granodiorite crops out in the Weeksboro-Lunksoos Lake anticline and a smaller pluton of pegmatitic granite crops out in the Liberty-Orrington anticline.

Plutons that clearly cut the Siluro-Devonian sequence are more abundant than the earlier plutons. Radiometric ages of the "granitic" plutons show considerable scatter, the cause of which is somewhat uncertain. If these plutons represent a single episode of intrusion, their age can be set approximately at 380 ± 25 my. Field observations show that rocks registering dates within this time range are of two ages - an older set of granites and graniodiorites that are intruded by younger granites and quartz monzonites.

Various mafic and ultramafic plutons are present in the coastal region (Fig. 1). These plutons range from serpentinites and pyroxenites to layered gabbro. Although some of these clearly intrude Lower Devonian rocks, it is not clear that they are all so young. Some are older than the felsic plutons of Devonian age. A few mafic dikes with chilled margins are thought to be of Triassic (?) age.

The metamorphic intensity increases regionally toward the south, and in contact aureoles it increases toward Devonian plutons. The contact aureoles contain andalusite and locally sillimanite in pelitic rocks and

calcsilicate minerals in carbonate-bearing rocks. The regional metamorphism may be due to the superposition of effects from several episodes of metamorphism. The sillimanite isograd shown in Figure 1 most likely expresses the summation of these effects. The rocks in the core of the Weeksboro-Lunksoos Lake anticline were recrystallized to low grade before deposition of Middle Ordovician rocks, and the Precambrian and Cambro-Ordovician rocks of east-central Maine were probably metamorphosed before the deposition of Siluro-Devonian rocks, although here the effects are difficult to sort out because of subsequent higher grade metamorphism. The Siluro-Devonian rocks along with the older rocks were subsequently recrystallized by Buchan-type metamorphism that is largely responsible for the regional distribution of isograds. This pervasive metamorphism is thought to post-date the Devonian plutonism on the basis of the decay of the K-Ar dates in the plutons located in metamorphic terrains of high grade, and on the basis that several plutons are cut by faults that do not displace the regional isograds. Plutons located in metamorphosed terrain of highest grade display K-Ar ages of 250 $\pm$ 10 my with high frequency, and these dates are interpreted as having been reset by the metamorphic event giving a maximum age of permian to the metamorphism.

Quaternary deposits form a nearly ubiquitos cover. The deposits in the coastal region were left during the recession of the Late Wisconsin ice sheet, between 13,500 and 12,300 years ago. These deposits include wide spread emerged marine deltas, strandline and/deeper water sediments, extensive systems of stratified moraines and small washboard moraines. The inland region is characterized by ground moraine, eskers, and other ice contact deposits, all overlain by a discontinuous cover of marine clays and silts.

The field trips deal with various aspects of the geology so broadly developed above. The reader will note that differing conclusions and interpretations have been reached in the course of the detailed work by different geologists, but these differences are a healthy sign that fruitful work is actively in progress. Additional work is still needed, including topical studies, to illucidate some of the questions and ambiguities which prevent deciphering the geologic history of Maine.

Philip H. Osberg
Editor

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Table of Contents

METAMORPHISM IN THE BELFAST AREA, MAINE

Charles E. Bickel
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Introduction

On this excursion we will consider regional metamorphism in the Brooks and Belfast 15 minute Quadrangles. This area lies near the northeastern terminus of high grade metamorphism in New England (Thompson and Norton, 1968). Five compositional groups of stratified rocks have been metamorphosed in the Belfast area:

1. Pelitic (quartzose-aluminous) rocks - normal pelite; siltstone or wacke without carbonate minerals
2. Calcic-aluminous rocks - limey pelite, limey siltstone, and limy wacke
3. Calcareous rocks - limestone and dolomitic limestone; the ratio of Mg:Fe is much greater in this group than in group 2
4. Mafic rocks - mafic and intermediate volcanic rocks
5. Quartzo-feldspathic rocks - felsic volcanic rocks; arkosic sediment

The metamorphism of rocks belonging to groups 1 and 2 is emphasized on this trip. Recrystallization of the rocks was contemporaneous with their folding, and occurred at low pressure. The age of metamorphism is discussed in the foreword of this guidebook. Some of the granites identified on Figure 1 are foliated and are contemporaneous with the regional metamorphism. Others are unfoliated and are interpreted to cut the regional isograds. Field studies of the area are nearly completed, but much petrographic and analytical work remains to be done.

Stratigraphic units referred to in the text and itinerary are described by: Bickel (1971), Bickel (in press), Cheney (1967), Doyle and Warner (1965), Ludman and Griffin (this guidebook), Osberg (1968), Osberg et al. (1968), Osberg and Guidotti (this guidebook), and Perkins and Smith (1925).

Acknowledgements

The Maine Geological Survey, under the direction of R.G. Doyle, supported my field work in the Belfast area during the summers of 1967-1973. The early work was supervised by Professors M.P. Billings and J.B. Thompson, Jr., of Harvard University. I gratefully acknowledge the financial support of Harvard from 1967 to 1971 and of the Frederic Burk Foundation for Education from 1972 to 1974. The electron microprobe analyses were made at the University of California, Berkeley.

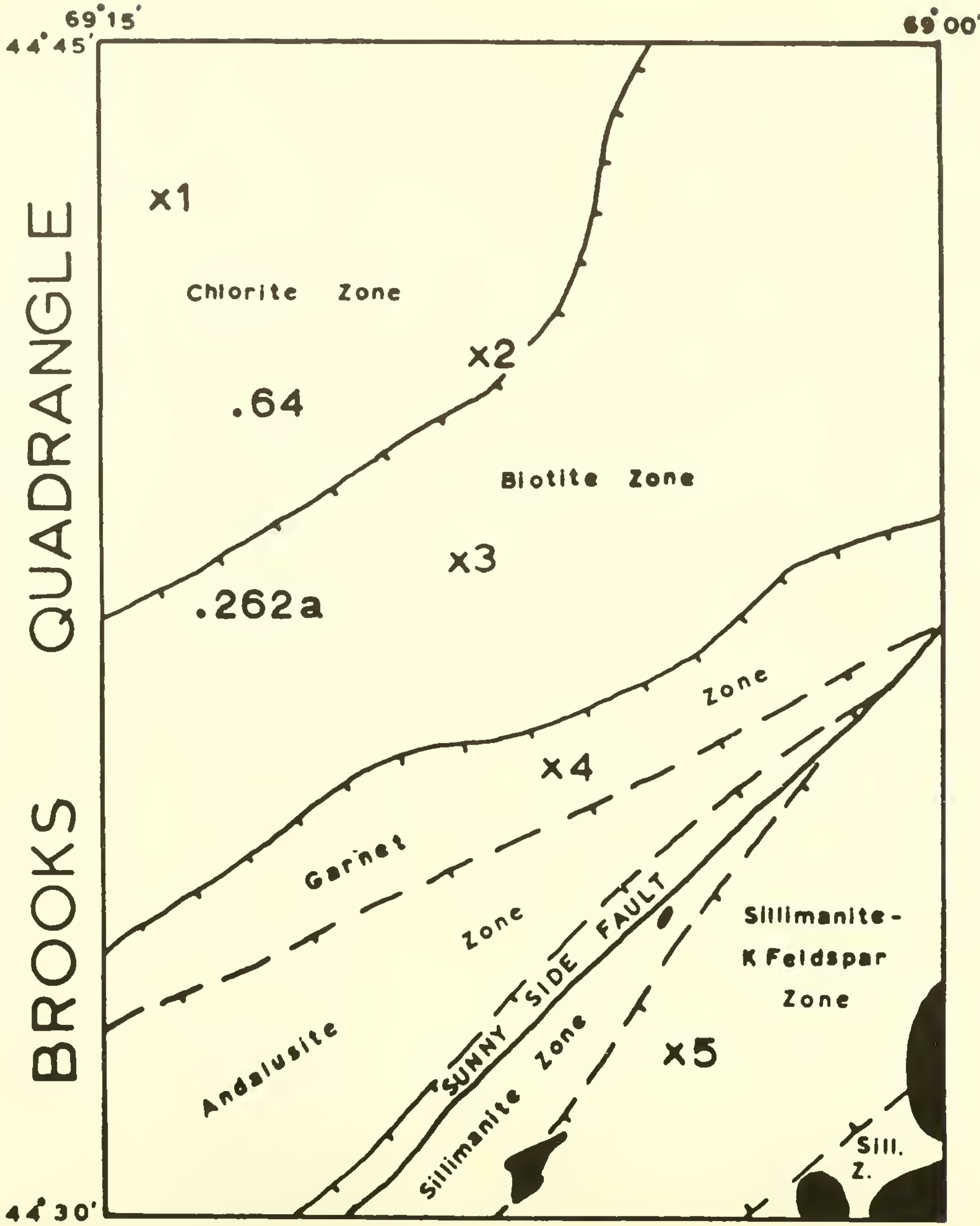
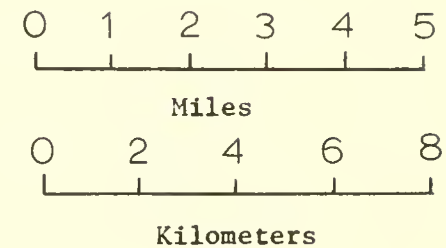
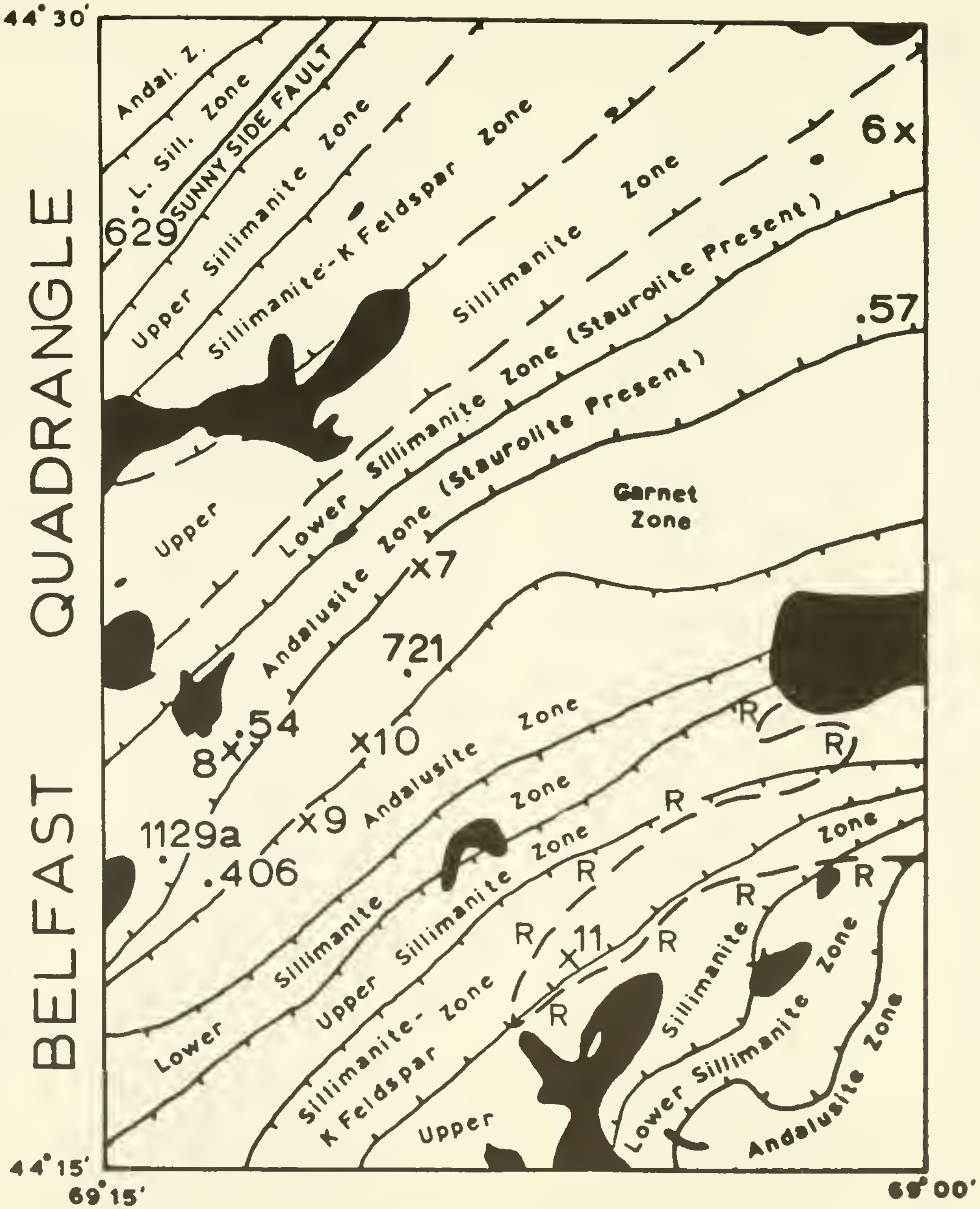


Figure 1. Distribution of the zones of regional metamorphism and of retrograde metamorphism in the Belfast area.



The isograds of regional metamorphism are hachured on the high-intensity side; dashes indicate that there is little control on the location of the isograd. The tracts of the andalusite zone and the lower sillimanite zone within which staurolite is found (in the Appleton Ridge Formation) are indicated.



The heavy dashed line labelled "R" approximately outlines the area in which there has been significant retrograde metamorphism (primarily replacement of biotite by chlorite and of aluminum silicate by sericite). The R's appear on the retrograded side of this boundary. The area is bounded on the northwest by the southwestern tract of the garnet zone.

indicates bodies of granitic rocks

x3 indicates a stop on our itinerary

.629 indicates the location of a mineral analysis referred to in the text

Distribution of the Zones of Prograde Metamorphism

Figure 1 shows the distribution of the metamorphic zones in the Brooks and Belfast Quadrangles. The zones are separated by isograds each of which marks the first appearance of an index mineral or mineral pair in rocks belonging to a particular compositional group. The andalusite and sillimanite isograds, together with the isograd separating the lower and upper sillimanite zone, relate to pelitic rocks. The sillimanite-K feldspar isograd marks the first appearance of that mineral pair in pelitic rocks or in aluminous quartzo-feldspathic gneisses of uncertain origin. Because pelitic rocks are rare in the central part of the Brooks Quadrangle, the garnet isograd necessarily is based on the first appearance of garnet in either pelitic or mafic rocks. The biotite isograd marks the first appearance of biotite in calcic-aluminous rocks because rocks of pelitic composition are not found in the northern part of the Brooks Quadrangle.

The isograds trend northeasterly, approximately parallel to the contacts of stratigraphic units. The lowest grade rocks of the area belong to the chlorite zone and lie in the northwestern part of the Brooks Quadrangle. The highest grade rocks belong to the sillimanite-K feldspar zone. They occur in two separate tracts, one of which lies in the central part of the area and one of which crosses the southeastern corner of the area. There is one tract of rocks in the biotite zone, but there are two separate tracts of rocks in the garnet zone and four separate tracts of rocks in the andalusite and sillimanite zones.

Data relating to the stratigraphy and metamorphism are confusing in the southern part of the Brooks Quadrangle, in the general vicinity of the Sunny Side Fault Zone. Possibly there are some lower grade rocks, bounded by faults, within the area mapped in the andalusite zone northwest of the fault line on Figure 1. The fault zone postdates the regional metamorphism, but the exact nature of offset of the isograds in this area is not presently known. John Griffin (personal communication, 1973) has noted a drop in metamorphic grade from the sillimanite-K feldspar zone, southeast of the fault, to the chlorite zone. This observation was made in the Bucksport Quadrangle, ten miles from where the fault leaves the Belfast area.

Prograde Metamorphism of the Stratified Rocks

Significant mineral assemblages of the calcic-aluminous, calcareous, and mafic compositional groups of rocks in each zone of prograde metamorphism are presented in Table 1. The mineral assemblages of pelitic rocks are represented graphically in Figure 2 (Thompson, 1957). All the compositional groups are not represented in all the metamorphic zones. Equations are presented below for certain metamorphic reactions suggested by petrographic evidence. The distinction between equations with and without numerical coefficients is explained by Thompson and Norton (1968, p. 321). The simplified mineral formulas used in most of the equations are those of Thompson and Norton, and many of the equations appear in their paper.

Table 1. Mineral Assemblages of Metamorphosed Calcic-aluminous Rocks, Calcareous Rocks, and Mafic and Intermediate Volcanic Rocks

	Chlorite Zone		Biotite Zone	Garnet Zone				Andalusite Zone									Sillimanite Zone													Sillimanite - K Feldspar Zone						
Compositional Group	2	2	2	2	4	4	4	2	2	3	3	3	3	4	4	4	2	2	2	2	2	3	3	3	4	4	4	4	4	2	2	2	2	3	3	4
Quartz	X	X	X	X	X	X	X	X	X	X	-	X	-	X	X	X	X	X	X	X	X	-	-	-	X	X	X	X	X	X	X	X	X	-	X	X
Albite	X	X	X	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Plagioclase	-	-	-	X	X	x	X	X	X	-	-	-	-	X	X	X	X	X	X	X	-	-	-	X	X	-	X	X	X	X	X	X	X	-	X	X
Microcline	-	-	-	-	-	-	-	X	-	-	-	-	-	-	-	-	X	-	X	-	-	-	-	X	-	-	-	-	X	-	X	-	-	-	X	-
Muscovite	X	X	X	X	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Biotite	-	-	X	X	X	X	X	X	X	-	-	-	-	X	X	X	X	X	-	-	-	-	-	-	X	X	X	X	x	X	X	-	X	-	-	-
Phlogopite	-	-	-	-	-	-	-	-	-	X	X	X	X	-	-	-	-	-	-	-	-	X	X	X	-	-	-	-	-	-	-	-	-	-	-	-
Chlorite	x	X	x	x	X	x	-	-	-	-	-	X	-	-	-	x	-	-	-	-	-	?	?	-	-	-	x	-	-	-	-	-	-	?	-	-
Garnet	-	-	-	-	-	X	X	-	-	-	-	-	-	-	X	X	-	-	-	X	X	x	-	-	X	X	-	-	-	-	-	X	-	x	-	-
Cordierite	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	X	-	-	-	-	-	-	-	-	-
Tremolite	-	-	-	-	-	-	-	-	-	-	X	X	X	-	-	-	-	-	-	-	-	-	X	X	-	-	-	-	-	-	-	-	-	-	-	-
Hornblende	-	-	-	-	-	x	x	-	-	-	-	-	-	X	X	X	-	-	-	-	-	-	-	-	-	X	-	X	X	-	-	-	-	-	-	X
Calcic amphibole #	-	-	-	-	-	-	-	X	X	-	-	-	-	-	-	-	X	X	X	-	X	-	-	-	-	-	-	-	-	X	X	X	X	-	X	-
Cummingtonite	-	-	-	-	-	X	-	-	-	-	-	-	-	-	-	X	-	-	-	-	-	-	-	-	X	X	-	X	-	-	-	-	-	-	-	-
Anthophyllite	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	X	-	-	-	-	-	-	-	-	-
Diopside	-	-	-	-	-	-	-	X	X	-	-	-	X	-	-	-	X	X	X	X	X	-	-	X	-	-	-	-	-	X	X	-	X	-	X	-
Ferrosalite *	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	X	-	-	-	-	-	-	X
Clinozoisite	-	-	-	-	X	-	x	-	X	-	-	-	X	X	-	-	-	X	X	X	X	-	-	-	-	-	-	-	-	-	-	-	X	-	X	-
Epidote	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	X	X	-	-	-	-	-	-	X
Forsterite	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	X	X	-	-	-	-	-	-	-	-	-	-	X	-	-
Humite Group	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	X	-	-	-	-	-	-	-	-	-	-	X	-	-
Sphene	-	-	-	-	-	-	x	X	X	-	-	-	X	X	X	-	X	X	X	X	X	-	-	X	-	-	-	X	X	X	X	X	X	-	X	X
Calcite	x	-	X	X	-	-	X	x	x	X	X	X	X	-	-	-	x	x	x	x	x	X	X	X	-	-	-	-	-	-	-	-	-	X	X	-
Dolomite	-	-	-	-	-	-	-	-	-	X	X	X	-	-	-	-	-	-	-	-	-	X	X	-	-	-	-	-	-	-	-	-	-	X	-	-
Ankerite	X	X	x	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Siderite	-	x	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Pleonaste	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	X	-	-	-	-	-	-	-	-	-

Explanation of Symbols:

Compositional groups of rocks are identified as in the text: 2 - calcic-aluminous; 3 - calcareous; 4 - mafic or intermediate volcanic.

X - consistently present

x - rare or not consistently present

? - magnesian chlorite; probably prograde

- - not present

- calcic amphibole denotes a calcium-bearing amphibole belonging to the tremolite-actinolite or to the hornblende series

* - ferrosalite is an iron-rich member of the diopside-hedenbergite series

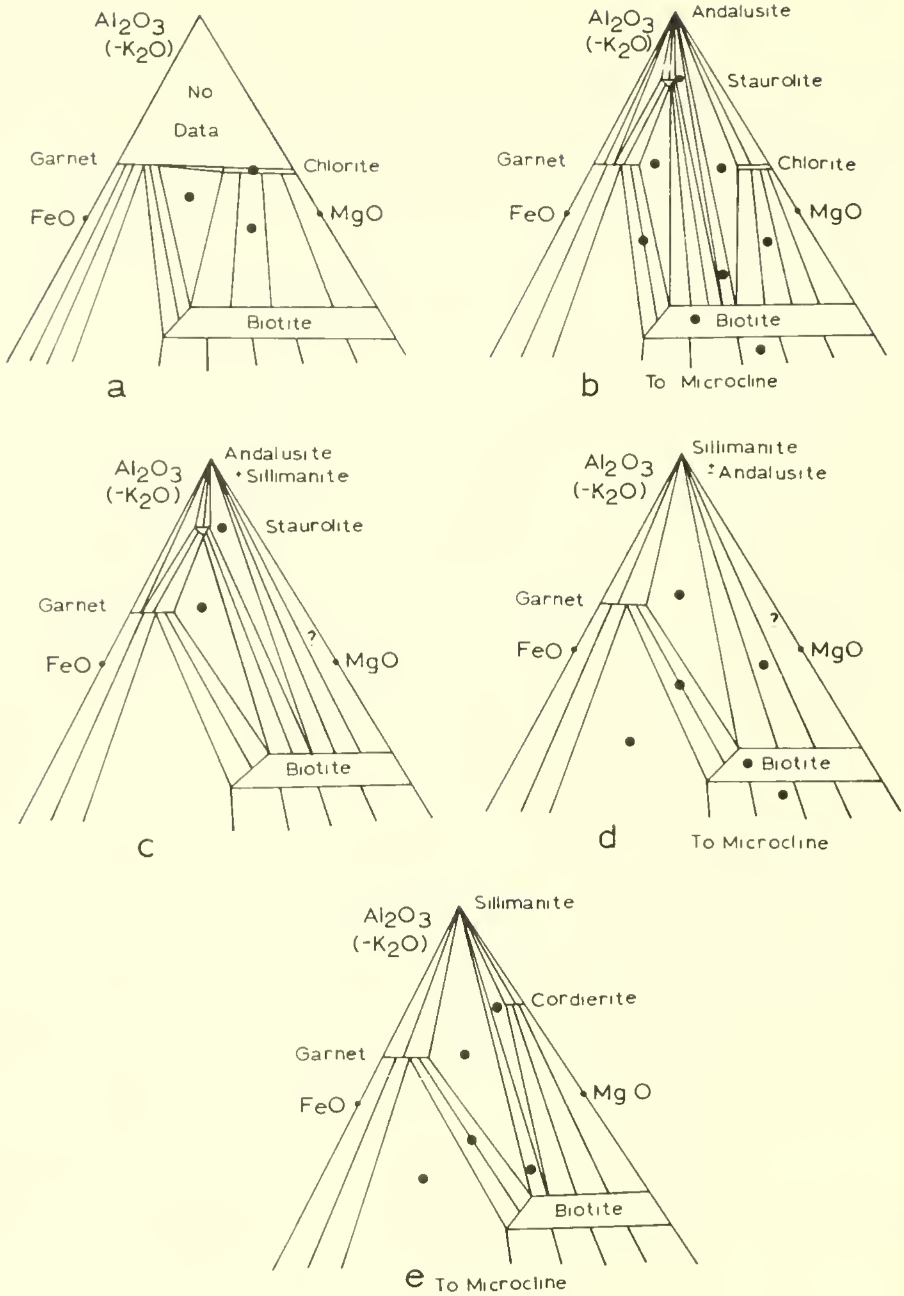


Figure 2. Diagrammatic Thompson projections of mineral assemblages with quartz and muscovite. Observed mineral assemblages are indicated: •

2a. Garnet zone. The rocks are not sufficiently aluminous for the development of minerals richer in aluminum than garnet or chlorite.
2b. Andalusite zone. The assemblage andalusite-staurolite-biotite always includes manganese-bearing garnet.
2c. Lower sillimanite zone (Appleton Ridge Formation). The assemblage andalusite-staurolite-biotite always includes manganese-bearing garnet.
2d. Lower and upper sillimanite zones, exclusive of the Appleton Ridge Formation. Andalusite generally is absent from the upper sillimanite zone.
2e. Sillimanite-K feldspar zone. Mineral assemblages that include sillimanite and K feldspar cannot be represented on this projection.

Chlorite zone. The slates and metamorphosed quartz wacke and siltstone of the chlorite zone contain minor or abundant carbonate minerals (Table 2). Siderite is abundant only in black slates, and perhaps is restricted to them. Chlorite is abundant in some slates, but in no other rocks. Sedimentary textures generally are preserved. They are modified chiefly by the development of slaty cleavage and carbonate porphyroblasts.

Table 2. Electron Microprobe Analyses of Carbonates from the Chlorite and Biotite Zones

	Chlorite		Zone		Biotite Zone
Location (Fig. 1)	64		Stop 2		262a
Mol %	calcite	ankerite	ankerite	siderite	ankerite
calcite	95.0	52.8	51.5	0.5	52.7
magnesite	1.9	30.4	31.5	42.9	33.3
siderite	2.5	15.8	16.5	56.5	13.3
rhodochrosite	0.6	1.0	0.5	0.1	0.7

Mineral assemblages and rock identifications:
64: quartz-albite-muscovite-pyrite; metamorphosed quartz wacke
Stop 2: quartz-albite-muscovite-chlorite; black slate
262a: quartz-albite-muscovite-chlorite-biotite-pyrite; metamorphosed greywacke

Biotite zone. The biotite isograd marks the first appearance of biotite. Limited data suggest that the mineral appears in quartz wacke at lower grade than in slate. In the lower biotite zone biotite generally is very fine-grained, comprising under one percent of the rocks, and therefore the biotite isograd cannot be detected in the field. Chlorite is much less abundant in the chlorite zone than is biotite in the middle and upper parts of the biotite zone, suggesting that biotite was produced by reaction of ankerite and siderite, as well as chlorite, with quartz and muscovite. Siderite is found only in slaty rocks near the biotite isograd, but ankerite (Table 2) and calcite occur as minor constituents of rocks throughout the biotite zone. The absence of amphiboles in the biotite zone deserves comment. In this area, high activity of CO_2 apparently inhibited the reaction:

$$5 \text{ ankerite} + 8 \text{ quartz} + H_2O \longrightarrow \text{actinolite} + 3 \text{ calcite} + 7\ CO_2 \quad (1)$$

High activity of CO_2 relative to H_2O probably is also responsible for the absence of epidote minerals. Relic sedimentary textural features are present only in the lower biotite zone. At higher grade in this zone the rocks are schistose and purplish due to the growth of megascopic biotite.

Garnet zone. The garnet "isograd" marks the first appearance of garnet in rocks of either pelitic or mafic composition. Garnet from both compositional groups of rocks is rich in the spessartite component (Table 3). In the pelitic rocks garnet and biotite probably have been produced by reaction between iron-rich chlorite, muscovite, and quartz. Mineral assemblages of pelitic rocks in the garnet zone are depicted in Figure 2a. Coarse sedimentary and volcanic clasts are preserved in rare beds in the garnet zone, but in general the rocks are coarse-grained phyllites or fine-grained schists with schistose or granoblastic textures.

Andalusite zone. The andalusite isograd marks the first appearance of andalusite in the pelitic rocks. In the Appleton Ridge Formation, which occupies the tract of andalusite-grade rocks distinguished on Figure 1, staurolite appears about at the andalusite isograd. It appears that the reactions:

$$\text{chlorite} + \text{muscovite} + \text{garnet} \longrightarrow \text{staurolite} + \text{biotite} + \text{quartz} + H_2O \quad (2)$$

$$\text{chlorite} + \text{muscovite} + \text{staurolite} + \text{quartz} \longrightarrow \text{andalusite} + \text{biotite} + H_2O \quad (3)$$

occurred at about the same conditions of P, T, and aH_2O. In more magnesian rocks the andalusite isograd probably marks the reaction:

$$3\ \text{chlorite} + 7\ \text{muscovite} + \text{quartz} \longrightarrow 13\ \text{andalusite} + 7\ \text{biotite} + 18\ H_2O \quad (4)$$

in which the chlorite is more magnesian than the chlorite of reactions (2) and (3). The mineral assemblages of pelitic rocks in this zone are represented in Figure 2b. The assemblages with staurolite are restricted to the Appleton Ridge Formation. There is a significant component of spessartite in the garnet of the common assemblage:

quartz-muscovite-biotite-garnet-staurolite-andalusite

(Table 3). Green (1963) and Osberg (1971) have discussed the occurrence and meaning of "extra phases" on Thompson projections.

Inconsistent mineral assemblages are indicated in Table 1 for marbles in the middle of the andalusite zone. In these marbles the reaction:

$$5\ \text{dolomite} + 8\ \text{quartz} + H_2O \longrightarrow \text{tremolite} + 3\ \text{calcite} + 7\ CO_2 \quad (5)$$

was dependent on the local relative activities of CO_2 and H_2O. The reaction:

$$\text{tremolite} + 3\ \text{calcite} + 2\ \text{quartz} \longrightarrow 5\ \text{diopside} + 3\ CO_2 + H_2O \quad (6)$$

was dependent on the local permeability of the rocks to escaping CO_2 and H_2O.

Pelitic rocks in the andalusite zone are moderately or strongly schistose. Primary textures are preserved only in beds of conglomerate and coarse tuff.

Table 3. Electron Microprobe Analyses of Garnets from the Garnet, Andalusite, and Lower Sillimanite Zones

Locality (Fig. 1)		Almandine	Spessartite	Grossularite	Pyrope	Vol. % Garnet	Mineral Assemblage, Protolith, and Formation
		Mol % **					
721	#	34.6	55.3	9.0	1.1	20	Qtz-Mus-Biot-Chl; Pelite, probably with minor mafic ash; Appleton Ridge Fm.
	#	48.5	42.8	6.8	1.9		
406	#	38.4	43.5	17.4	0.7	35	Qtz-Cum-Hbd-Biot; Mafic tuff; Muzzy Ridge Lentil of Appleton Ridge Fm.
	#	47.7	41.7	9.9	0.7		
57	core	76.4	13.4	3.4	6.8	< 1	Qtz-Mus-Biot-Staur-Andal-Chl*; Pelite; Appleton Ridge Fm.
	rim	78.6	12.5	2.2	6.7		
54	core	78.0	10.8	3.7	7.5	2	Qtz-Mus-Biot-Staur-Chl*; Pelite; Appleton Ridge Fm.
	rim	80.1	9.3	2.6	8.0		
1129a	core	69.4	22.8	4.3	3.5	3	Qtz-Mus-Biot-Staur-Andal-Chl*; Pelite; Appleton Ridge Fm.
	rim	83.1	6.1	4.4	6.4		
629	core	63.2	24.9	3.6	8.3	3	Qtz-Calcic Olig-Mus-Biot-Andal-Fibrolitic Sill; Graywacke; Hogback Schist
	rim	68.4	18.9	3.2	9.5		

** All iron is assumed to be FeO
# indicates an extreme composition of garnet that is irregularly zoned
Chl* indicates iron-rich chlorite that is believed to be retrograde
Locations 721 and 406 are in the garnet zone; locations 57, 54, and 1129a are in the andalusite zone; location 629 is in the lower sillimanite zone

Lower sillimanite zone. The sillimanite isograd marks the first appearance of sillimanite in the pelitic rocks. It appears first as fibrolitic sillimanite that is generally too rare and fine-grained to detect in hand specimens. It is likely that most of the fibrolitic sillimanite was produced by the reaction:

$$\text{almandine} + \text{muscovite} \longrightarrow 2\ \text{sillimanite} + \text{biotite} + \text{quartz} \quad (7)$$

Fibrolitic sillimanite may have been produced in some of the rocks by reactions such as:

$$6\ \text{staurolite} + 4\ \text{muscovite} + 7\ \text{quartz} \longrightarrow 31\ \text{sillimanite} + 4\ \text{biotite} + 3\ H_2O \quad (8)$$

$$\text{staurolite} + \text{muscovite} + \text{quartz} \longrightarrow \text{sillimanite} + \text{almandine} + H_2O \quad (9)$$

Andalusite survives the first appearance of sillimanite and is in fact the predominant aluminum silicate throughout the lower sillimanite zone. The presence of andalusite and the almost complete absence of coarse sillimanite distinguish the lower from the upper sillimanite zone.

The mineral assemblages of pelitic rocks in the lower sillimanite zone are represented in Figures 2c and 2d. The restriction of staurolite to the Appleton Ridge Formation in both the andalusite and the lower sillimanite zone is probably due to the low manganese and calcium content of those rocks in comparison with the other schists of the area. Analysed garnet from the Hogback Schist (629 in Table 3) has a greater spessartite content than does garnet from the Appleton Ridge Formation (57, 54, 1129a in Table 3). In rocks such as the Hogback Schist, the assemblage (with quartz and muscovite):

biotite-almandine(Mn-rich)-andalusite±sillimanite

is stable relative to the assemblage:

biotite-almandine(Mn-poor)-staurolite±andalusite±sillimanite

of the Appleton Ridge Formation. Thus the two configurations of Thompson projections for the lower sillimanite zone in Figures 2c and 2d are not necessarily the result of an increase in metamorphic grade from Figure 2c to Figure 2d.

Pelitic rocks in the lower sillimanite zone are schistose and appear similar to those of the andalusite zone.

Upper sillimanite zone. The lower and upper sillimanite zones are separated by an isograd that marks the appearance of abundant coarse sillimanite in the pelitic rocks. Insufficient data for the subdivision of the sillimanite zone exists in the Brooks Quadrangle. Because andalusite is rare in the

upper sillimanite zone, it appears that the coarse sillimanite formed by inversion of andalusite. Fibrolitic sillimanite is common in the pelitic rocks of the upper sillimanite zone. Figure 2d represents the mineral assemblages of these rocks. Their textures are granoblastic or gneissic.

Sillimanite-K feldspar zone. The sillimanite-K feldspar isograd encompasses all localities where sillimanite and microcline occur together. It was necessary to study thin sections to locate the isograd. All of the specimens with microcline and sillimanite also contain prograde muscovite. Microcline has been produced in the sillimanite-bearing rocks of this zone by the reaction:

$$\text{muscovite} + \text{Na-plagioclase} + \text{quartz} \longrightarrow \text{sillimanite} + \text{K-Na-feldspar} + H_2O \quad (10)$$

This reaction is dependent on the bulk composition of the rocks and has been studied by Evans and Guidotti (1966). Fibrolitic sillimanite is common in this zone. In one specimen there is andalusite together with coarse sillimanite. Mineral assemblages of these rocks are represented in Figure 2e. Cordierite first appears in pelitic rocks in the sillimanite-K feldspar zone. The cordierite is restricted to sulfidic rocks in which iron is incorporated in pyrrhotite or pyrite. The bulk composition of these rocks is effectively enriched in magnesium. The pelitic rocks of the sillimanite-K feldspar zone have granoblastic or gneissic textures.

Inferences about the Rate of Heating During Prograde Metamorphism

It is likely that the prograde metamorphism was caused by rapid heating. Probable disequilibrium not related to retrograde metamorphism characterizes many of the rocks. In particular, andalusite predominates in the lower sillimanite zone and persists into the sillimanite-K feldspar zone. The garnets of Table 3 are both regularly and irregularly zoned. Electron microprobe analyses indicate that pyroxenes and particularly amphiboles of metamorphic origin are not homogeneous in their distribution of Si-Al and Mg-Fe. Furthermore, the primary igneous minerals of metamorphosed mafic intrusive rocks (Bickel, 1971) are by far the best preserved in the sillimanite-K feldspar zone. This would be unlikely if these rocks had passed slowly through conditions of the lower metamorphic zones.

Retrograde Metamorphism

In the Belfast Quadrangle some of the prograde metamorphic minerals have been replaced by hydrous minerals characteristic of lower metamorphic grades.

Generally such replacements have occurred only southeast of the southeastern tract of the garnet zone. The area of most intense retrograde metamorphism is outlined on Figure 1. This event has affected a majority of the pelitic rocks and many of the mafic rocks, but has had little effect on marbles, calc-silicate granofelses, and biotite gneisses. The nature of mineral replacement during retrograde metamorphism is summarized in Table 4. Oligoclase or more calcic plagioclase is not replaced by albite and clinozoisite. The assemblage quartz-muscovite-chlorite-microcline, representative of the chlorite zone, was not produced by the retrograde metamorphism. These facts, together with the spatial relationship between chloritization of biotite and sericitization of aluminum silicate in retrograded pelitic rocks suggest that retrograde metamorphism occurred under conditions of the garnet or biotite zone. Garnet has been only sparingly replaced by chlorite in the retrograded pelitic rocks. Perhaps the replacements took place in the biotite zone and garnet was preserved due to kinetic constraints.

Retrograde metamorphism may have occurred during cooling or reheating of the rocks following the major prograde metamorphic events. It consisted of local, partial re-equilibration , at lower temperature, of the high-grade rocks in the southern part of the area. It was promoted by the local re-introduction of considerable aqueous fluid.

Table 4. Mineral Replacements of Retrograde Metamorphism

Rock Compositional Group	Original Mineral #		Replacement Mineral *	Comments
1. Pelitic	andalusite or sillimanite	→	sericitic muscovite + quartz	
	biotite	→	chlorite	
	andalusite or sillimanite	→	sericitic muscovite + chloritoid + quartz	See description of Stop 9
	biotite	→	chlorite	
	staurolite	→	iron-rich chlorite + muscovite	
	biotite	→	chlorite	
	garnet	→	chlorite or muscovite + quartz	This replacement is much less common and complete than the replacements listed above
2. Calcic-aluminous	biotite	→	chlorite	This replacement is rare
3. Calcareous	forsterite	→	serpentine	
	? ? ?	→	chlorite	
	? ? ?	→	tremolite	
4. Mafic	anthophyllite	→	talc	
	garnet	→	chlorite	
	? ? ?	→	serpentine	
5. Quartzo-feldspathic	biotite	→	chlorite	This replacement is rare

- replacements enclosed by brackets are spatially related
* - the texture of the replacement minerals is decussate

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Itinerary

Mileage	
0	Assemble in the center of Plymouth, with cars facing south on Route 7 at its intersection with Route 69. The assembly point is across from the Village Store and Post Office. Starting time 9:00 A.M. Proceed south on Route 7.
200 ft.	Turn right on hardtop road with no route number. Cross Martin Stream, which drains Plymouth Pond.
0.4	Continue straight, following sign for Round Pond.
1.4	Enter Brooks Quadrangle.
1.9	At crossroads turn left on unpaved road, again following sign for Round Pond.
2.6	Round Pond is to the left. Continue straight.
3.6	Enter Waldo County (no road sign).
4.2	<u>Stop 1</u>. Park along the road, beyond the white house with green trim. To the left are polished pavement outcrops of the Kenduskeag Formation in the chlorite zone. NO HAMMERING IN AREAS WITH POLISHED SURFACES, PLEASE. The northern part of the outcrop is thin-bedded slate. Tops are to the southeast. South of a 75 foot covered section is a whitish-weathering unit, predominately of slate, that appears massive. Here bedding features have been nearly obliterated by shearing. Further south is unpolished pavement of thin-bedded slate. The mineral assemblage quartz-albite-muscovite-chlorite-ankerite characterizes all of these rocks. Open folds of the bedding and slaty cleavage are associated with a slip cleavage. Veins of quartz, calcite, and minor ankerite are parallel to this slip cleavage. Return to cars promptly. Continue south.
4.5	Cross power lines.
6.1	Road becomes paved.
6.2	Continue straight on paved road.
6.6	Continue straight (south) through crossroads. Road becomes unpaved.
7.7	Road becomes paved.

8.3	Turn left on Route 9 and U.S. Route 202 at Troy. Continue east on Route 202 until Stop 2.
8.5 to 9.1	Good outcrops of the Vassalboro Formation in the chlorite zone.
9.8	Good outcrop of Kenduskeag Formation in the chlorite zone.
11.0	Enter Penobscot County.
12.6	Cross Route 7 at Dixmont.
14.1	Stop 2. Park at Dixmont Town House, the red building on the right, just before the power lines. Walk about 150 feet west on Route 202 to the second woods road running north. Follow this road about 200 feet to a pavement outcrop near the power lines. Graded beds of quartz wacke and slate characteristic of the Vassalboro Formation top to the northwest. These rocks are in the chlorite zone near the biotite isograd. Follow the power lines back to Route 202, passing more outcrops of quartz wacke. Walk 350 feet east to the road cut at the junction of Route 202 and Garland Road. Here quartz wacke alternates with thin-bedded black slate. In addition to quartz, albite, chlorite, and muscovite, the slate contains porphyroblasts of ankerite and siderite that are flattened in the plane of the slaty cleavage. The coarse, brown-weathering spots are aggregates of quartz, ankerite, siderite, muscovite, and chlorite separated from the slaty matrix of the rock by a thin mantle of siderite crystals. Analyses of the ankerite and siderite in this slate appear in Table 2. The quartz wacke contains quartz, albite, muscovite, ankerite, and possibly rare siderite. Return to cars. Continue east on Route 202 to Garland Road.
14.2	Turn right on unpaved Garland Road.
14.6	Continue straight on improved road.
14.8	Cross power lines.
15.6	Turn right at crossroads.
16.2	Cross power lines.
16.7	Bear right (south) at road junction.
17.4	Road curves sharply right.
18.3	Road curves sharply left.

18.8	Road becomes paved.
19.6	Bear left on paved road and slow down.
19.7	Stop 3. Park along road. We will study outcrops along the road from the granddaddy pine to the road junction. The Vassalboro Formation (?) is here in the middle of the biotite zone. Biotite generally is more abundant than muscovite in these rocks. Folded laminations in the quartzose beds are lithologic layers that developed parallel to an early cleavage. They are easily confused with bedding features. Biotite lies in a schistosity parallel to the axial planes of the folds. In beds of mica schist the biotite forms conspicuous porphyroblasts. There are a few punky-weathering beds that contain calcite. Carbonate minerals are less abundant in these outcrops than is typical of the biotite zone. Megascopic ankerite is restricted to quartz veins, but rare, microscopic ankerite is present in some of the mica schist. The opaque minerals are magnetite and pyrrhotite. Return to cars. Proceed south.
21.3	Cross brook.
21.5	Road ends. Turn left (east) towards Jackson.
21.7	Turn right (south) on unpaved road. You are in Jackson.
22.4	Road becomes paved.
23.5	Stop 4. (Optional stop) Park along the road beside a dark outcrop at the edge of the woods east of the road. This black, garnetiferous schist belongs to an unnamed, thin unit of metavolcanic rocks in the garnet zone. In addition to abundant biotite and garnet, the rock contains quartz, plagioclase, calcite, rare hornblende, probable clinozoisite, and what appears to be metamict sphene. Return to cars. Proceed south.
25.1	Proceed straight across Route 139 onto unpaved road.
25.7	Cross stream.
26.0	Turn right on paved road.
26.4	Cross railroad.
26.8	Turn very sharp left onto unpaved road and descend hill.
27.4	Continue straight.

27.6 Cross railroad.

28.8 Follow main road curving to right.

29.7 Stop 5. Park along road, leaving room for infrequent traffic. Pavement outcrops are to the left of the road. The Passagassawakeag (Knox) Gneiss is in the northwestern tract of the sillimanite-K feldspar zone. Lithologic layering interpreted to be bedding is much better preserved in these outcrops than is typical of the Passagassawakeag Gneiss. Most beds essentially consist of quartz, plagioclase, and biotite. Slightly more aluminous beds also contain minor muscovite, fibrolitic sillimanite, and microcline. Muscovite has incompletely reacted with quartz and plagioclase to produce sillimanite and microcline. The microcline forms porphyroblasts in quartz-plagioclase layers closely associated with the sillimanite, which is intergrown with muscovite and biotite. The core of a southwest-plunging, isoclinal synform is occupied by an unusual bed of hornblende-garnet granofels. This rock contains sphene and abundant calcic plagioclase and quartz. The foliation is folded here and elsewhere in the outcrop. Micas have not recrystallized parallel to the axial planes of these folds. NO HAMMERING IN AREAS WITH STRUCTURES, PLEASE. Weakly foliated granitic rocks (quartz-plagioclase-biotite) form pods. Augen gneiss is rare here, but common elsewhere in the Passagassawakeag Gneiss.
Return to cars. Proceed south.

31.1 Turn left (northeast) on Route 131. You are in the town of Waldo.

31.5 to 31.9 These outcrops together display the entire range of lithology and texture of the Passagassawakeag Gneiss in the sillimanite-K feldspar zone.

34.1 Follow Route 131 to the right.

35.4 Continue through crossroad on Route 131.

36.7 Bear right (south) on Route 141. Swan Lake is to the east. Continue south on Route 141 to Stop 6.

38.4 Enter Belfast 7½ and 15 minute Quadrangles.

39.4 Outcrops of Bucksport Formation and pegmatite to the right.

39.8 Stop 6. Park along highway, north of the Belfast City Boundary.

To the east are spectacular road cuts of the Bucksport Formation in the lower sillimanite zone. Lithologic layering here and elsewhere in the Bucksport Formation is much less strongly folded than is bedding of other rocks in the Belfast area. For this reason, some geologists believe that the lithologic layering here is not bedding, but is a compositional layering parallel to the axial planes of isoclinal folds. Beds (?) of biotite granofels (55 %), calc-silicate granofels (44 %), and biotite schist (1 %) are present. Most of the biotite schist is sulfidic and weathers rusty. The long axes of calcic amphibole plunge gently to the southeast. The minerals quartz, andesine or labradorite, microcline, biotite, calcic amphibole, diopside, clinozoisite, calcite, and sphene are all found in the Bucksport Formation in the andalusite and the sillimanite zones. Calcite and clinozoisite are not abundant, however, and generally calcite and diopside are absent from layers with biotite and microcline. The rocks of the Bucksport Formation belong to the calcic-aluminous compositional group, and thus are approximately the high-grade equivalents of the rocks we saw at Stops 1, 2, and 3. Both the granitic and pegmatitic dikes have a pinch-and-swell structure. The granitic dikes are concordant and are unfoliated or very weakly foliated. The pegmatitic dikes are unfoliated, discordant, and they cut the granitic dikes. The largest granitic body contains an irregularly-shaped pegmatitic phase within which there are graphic intergrowths of quartz and black tourmaline.
Return to cars. Continue south.

40.1 Another good outcrop of the Bucksport Formation.

43.5 Turn right on Route 3 and U.S. 1, crossing the Passagassawakeag River and by-passing Belfast.

44.9 Turn right on Route 3, heading west.

49.1 Enter Belmont.

49.4 Outcrop of muscovite-chlorite-biotite phyllite belonging to the Appleton Ridge Formation in the garnet zone.

50.5 Enter Morrill 7½ minute Quadrangle.

51.2 Turn left at Belmont Corner on Route 131, heading southwest.

52.5 Stop 7. Park along road beside outcrop (on right) of bedded phyllite and micaceous quartzite.
The Appleton Ridge Formation here is exposed in the garnet zone,

but garnet is rare in this outcrop and in most of the Appleton Ridge Formation at this grade. The rocks are composed primarily of quartz, muscovite, and chlorite with rare porphyroblasts of biotite that give them a spotted appearance. Graded beds suggest that tops are to the northwest. The nearly vertical schistosity is crenulated.
Return to cars. Continue southwest. Stay on Route 131 until mile 56.1.

53.4 Enter Searsmont $7\frac{1}{2}$ minute Quadrangle.

54.6 Continue straight, joining Route 173.

55.5 Cross St. George River and enter Searsmont.

55.6 Continue straight (southwest) uphill on Route 131, leaving Route 173.

56.1 Leave Route 131. Continue straight uphill on unpaved Appleton Ridge Scenic Drive. You will pass numerous outcrops of staurolite schist belonging to the Appleton Ridge Formation in the andalusite zone.

57.1 Stop 8. Park along the road about 200 feet beyond the hill crest and new house. Pavement outcrops of the Appleton Ridge Formation in the andalusite zone are to the right of the road. There is nothing of interest to geologists in the BLUEBERRY FIELDS; PLEASE KEEP OUT of them.
There are no reliable indications of the topping direction of the bedding. Staurolite schist is interbedded with micaceous quartzite. Staurolite, garnet, and biotite porphyroblasts are conspicuous. Tourmaline and poikiloblastic andalusite are much more difficult to discern. Biotite and staurolite have been partly replaced by iron-rich chlorite during retrograde metamorphism. Many of the quartz veins and pods contain pink andalusite. Staurolite is concentrated at the borders of some of the quartz veins. The schistosity is crenulated.
Return to cars. Continue southwest.

57.3 Turn left and descend the ridge.

57.7 Continue straight across Route 131 and across the St. George River at Ghent, where there are phyllites of the Appleton Ridge Formation in the garnet zone.

58.9 Turn left.

59.2 Turn right at Knights Corner on Moody Mountain road.

59.6 Do not confuse this first outcrop on the left with our next stop.

59.8 Stop 9. Park along road, beside an inconspicuous, gray outcrop to the left.
This spotted schist of the Jam Brook Formation belongs to the andalusite zone, but it has been extensively retrograded. The spots are decussate aggregates of sericitic muscovite, quartz, tabular chloritoid, and minor chlorite. The chloritoid tends to be concentrated at the borders of the spots. These minerals have almost completely replaced andalusite, which survives in the cores of some spots. Biotite has been partly or completely chloritized. Garnet has ragged grain boundaries; it may have been partly replaced by quartz and muscovite. But there has been very little, if any replacement of garnet by chlorite. Quartz veins are oriented parallel to a slip cleavage that is older than the retrograde metamorphism of the rock.

Chloritoid is absent from the prograde metamorphic rocks of the Belfast area. It was produced during retrograde metamorphism only where there were local compositions high in alumina. These locations are the sites of aluminum silicate minerals that grew during prograde metamorphism when chlorite and muscovite (±garnet) reacted to give aluminum-rich minerals and biotite. Compare the topology of Figure 2a with Figures 2b, 2c, 2d, and 2e.
Return to cars. Turn around and proceed north. (Road log assumes that you turned around at the outcrop.)

60.3 Continue straight on Moody Mountain Road.

62.1 Turn sharp right on Muzzy Ridge Road. (This turn is before Moody Mountain Road reaches Route 131, a few hundred feet to the northwest.) Proceed southeast up Muzzy Ridge.

63.5 Stop 10. (Optional stop) Park along road beyond ridge crest beside dark, ribbed outcrop under telephone pole to the left of road. NO HAMMERS.
These metamorphosed mafic volcanic rocks belong to the Muzzy Ridge Lentil of the Appleton Ridge Formation. They are weakly schistose and are exposed in the garnet zone. They are composed primarily of cummingtonite and garnet (similar to analysis 406 in Table 3), but also contain hornblende, quartz, plagioclase, biotite partly replaced by chlorite, graphite, and pyrite. The northeast-plunging folds end abruptly at shear planes.
Return to cars. Continue driving to the southeast.

63.7	Continue straight. Levenseller Mountain, to the southeast, is underlain by retrograded andalusite schist. Moody Mountain, to the south, is underlain by retrograded sillimanite gneiss.
65.4	Road curves sharply left.
65.8	Turn right on Route 173. Continue on Route 173 to mile 68.7.
66.2	Levenseller Pond is to the right. The area around the pond is underlain by granitic rocks.
66.6	Continue straight on Route 173.
66.9	Enter Lincolnville $7\frac{1}{2}$ minute Quadrangle.
68.7	Lincolnville Center. Continue straight.
68.8	Turn right on Route 235 and slow down.
68.9	Stop 11. Park along road beside rusty-weathering outcrop on right, from which a telephone pole has sprouted, bringing up fresh rock. Pyrrhotitic gneiss of the Penobscot Formation is in the sillimanite-K feldspar zone. Both fibrolitic and imperfectly oriented coarse sillimanite are abundant. The rock also contains muscovite, microcline, biotite, quartz, and plagioclase. Garnet is rare in the Penobscot Formation at all metamorphic grades because in these rocks considerable iron resides in pyrrhotite or pyrite.

Return to Orono via Route 52, which runs from Lincolnville Center to Belfast.

RECESSION OF THE LATE WISCONSIN LAURENTIDE ICE SHEET IN EASTERN MAINE

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Introduction

The fluctuating margin of the Late Wisconsin Laurentide Ice Sheet in Maine retreated approximately parallel to the coast leaving a belt of submarine end moraines (Borns, 1966, 1973). This recession was accompanied by a marine transgression of the coastal region that extended into the river valleys of central Maine (Goldthwait, 1949). Glaciomarine sediments up to approximately 150 feet thick were deposited in the coastal region. Inland of the end moraine belt central Maine is characterized by ground moraine, eskers up to 80 miles long extending to the southeast-facing slopes of the northeast-southwest-trending highlands of the state (Leavitt and Perkins, 1936) and glaciomarine sediments in the major river valleys.

The glacial geology of eastern coastal Maine is characterized by a northeast-southwest-trending 25 mile wide complex composed of hundreds of end moraines and associated features (Leavitt and Perkins, 1935; Borns 1966, 1967, 1973) deposited along a fluctuating glacier margin as it retreated northwest from a position on the continental shelf. Most, and perhaps all, of these moraines were deposited below the sea level that prevailed at that time. Within the end-moraine complex as many as 20 local marginal fluctuations are recognized.

End-Moraine Complex

The end-moraine complex is characterized by hundreds of end moraines, but also includes ice-marginal kames and marine deltas and interlobate deposits (fig. 1).

Two types of end moraines are recognized; a large, stratified and relatively continuous type in contrast to the mor numerous, small non-stratified and discontinuous "washboard" type. The large moraines are often up to 60 feet high, 300 feet wide with segments continuous for up to 10 miles in length. Internally these are composed predominately of stratified sand and gravel with minor interbeds of compact till and fossiliferous marine silts. Commonly these deposits have been deformed by ice push from their proximal sides which, coupled with the cross-cutting lobate map-pattern of the moraines, indicates that the moraines formed along the margin of an internally active ice sheet. These moraines are very similar in composition and origin to the Ra and Central Swedish moraines of Scandanavia.

The majority of the smaller and more numerous type of moraine seldom exceed 10 feet in height, 30 feet in width and 0.5 mile in length. Commonly

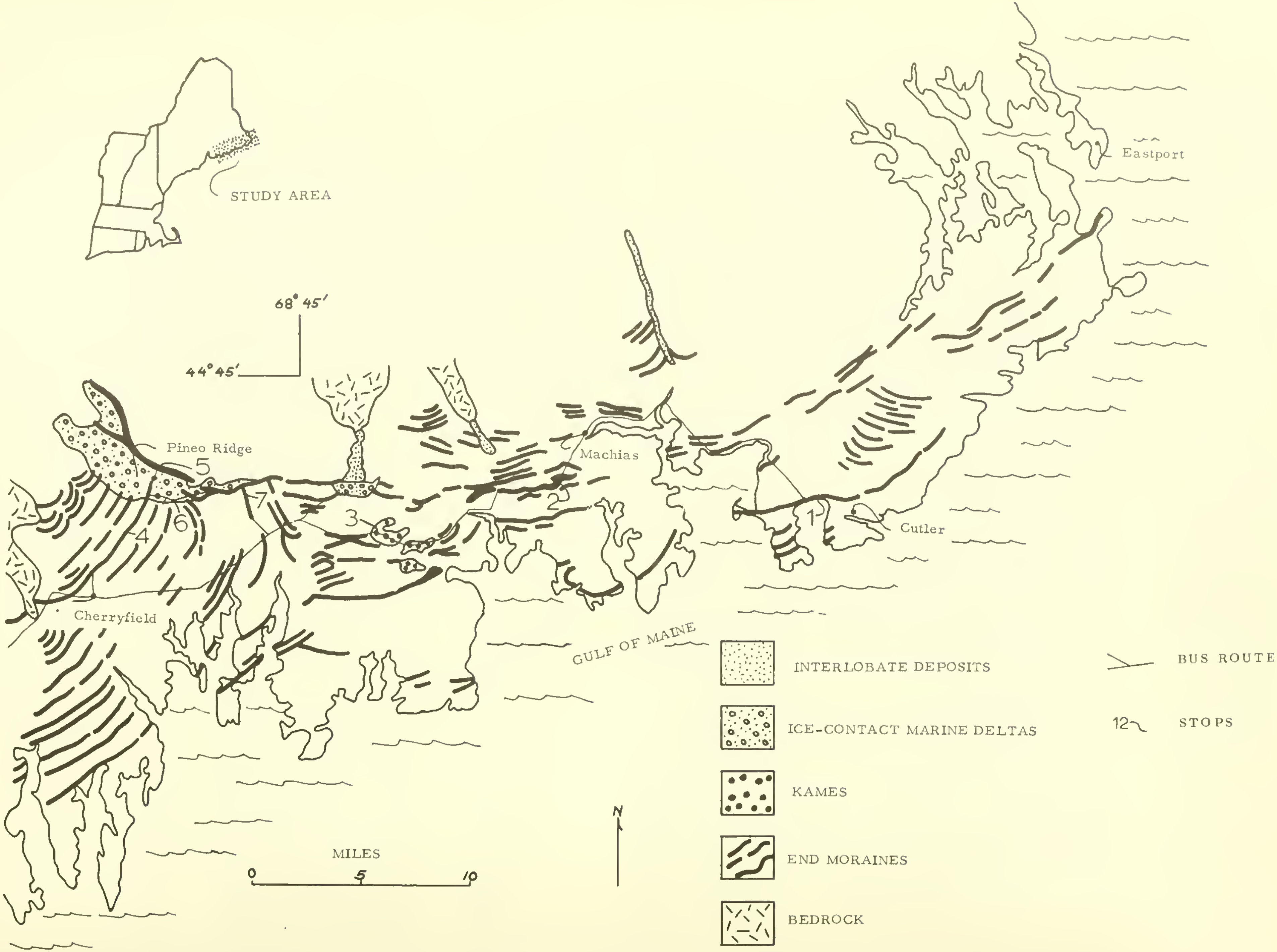

Figure 1. Generalized map showing the glacial geology of eastern Maine.

they occur in clusters of up to 50 parallel curved and evenly spaced moraines and hence the name "washboard" moraines. Nearly all of these moraines are composed of compact till.

The spatial and stratigraphic relationships of both types of moraines indicate that the larger stratified moraines formed during local readvances of the ice margin and that the smaller "washboard" moraines formed during the subsequent recessions. All the moraines found are below the upper marine limit and most probably formed below sea level. The processes responsible for the differences in the characteristics of these two types of moraines are now being studied.

The large volume of stratified drift within the large moraines and their wide distribution indicate that extensive melting of the ice sheet was prevalent and that while generally meltwater was discharging all along the ice margin local drainage concentrated large deposits of stratified drift as kames or marine deltas at the margin formed.

Marine Transgression

The recession of the ice sheet in the coastal and central sections of Maine was contemporaneous with a marine transgression. Evidence of both submergence and emergence is generally documented by the fossilferous silty clay deposits of the region which form a discontinuous cover partially filling the valleys and lapping up on the highlands to the altitude of the maximum postglacial marine submergence (Goldthwait, 1949). The marine sediment, named the Presumpscot Formation (Bloom, 1960), was deposited in the proximity of the receeding ice margin as indicated by the ice-marginal deltas, intertongued marine sediments within the end moraines, the abundance of ice-rafted erratics, and by the cold water marine fauna within the sediment.

Pineo Ridge Readvance

The glacier recession that produced the coastal moraine complex was interrupted by an extensive readvance in eastern Maine that terminated in the sea at Pineo Ridge Moraine approximately 12,800 to 12,600 years ago (Borns, 1973). Reconaissance of the distribution of deposits and of ice movement indicators demonstrate that the margin receded a minimum of 50 miles before readvancing to the position of Pineo Ridge Moraine.

Chronology

The Late Wisconsin terminal position of the Laurentide Ice Sheet east of the Hudson River is marked by the Ronkonkoma-Vineyard-Nantucket moraine line (Schafer, 1961; Kaye 1964; Schafer and Hartshorn, 1965) and probably by the distribution of coarse gravel on the continental shelf to the east (Schlee and Pratt, 1970).

The ice sheet reached its maximum extent in southeastern New England at least sometime after 20,000 years B.P. (Schafer and Hartshorn, 1965) and was probably still in this position on Marthas Vinyard, Massachusetts as late as 15,300 years B.P. (Kay 1964). By approximately 13,500 years B.P. the margin of the ice sheet had receded to the present position of the Maine coast and at least by 12,300 years B.P. the entire coastal moraine complex had been deposited (Stuiver and Borns, unpub. data).

Approximately 30 C^{14} dates on marine organisms contained in the emerged glaciomarine sediments and on the oldest organic sediments in lakes below the upper marine limit closely bracket the time of the marine submergence and therefore the formation of end moraines of the coastal region between 13,500 and 12,300 years B.P. (Borns, 1967; Stuiver and Borns, 1967; Stuiver and Borns, unpub. data).

Correlation

The deglaciation of coastal Maine and probably New Brunswick was well underway by 13,500 years B.P. This recession was interrupted by a readvance in the St. John River lowland of Maine and New Brunswick that terminated at the Pineo Ridge Moraine in eastern Maine about 12,700 years B.P.

In the Great Lakes region the Cary-Port Huron recession is documented by an intertill bryophyte bed in northern Michigan dated at 12,500 - 13,000 years B.P. and by low-level Lake Arkona in the Erie basin. The Port Huron readvance is recorded by till overlying the bryophyte bed, by the transition from Lake Arkona to Lake Whittlesey about 13,000 years B.P., and by the Port Huron Moraine.

The events in coastal Maine and the Great Lakes region are similar within the limits of C^{14} dates and stratigraphic control. Therefore Borns and Denton (1972) have suggested that the concept of the Cary-Port Huron recession and subsequent Port Huron readvance can be extended to eastern North America.

Rapid deglaciation of much of eastern North America and the Great Lakes area 13,500-13,000 years B.P. was followed by wide-spread readvance 13,000-12,500 years B.P. that culminated at the Port Huron Moraine and the Pineo Ridge Moraine. If this correlation proves to be correct, the widespread nature nature of these events would suggest an important climatic reversal that affected a large segment of the southern margin of the Laurentide Ice Sheet.

Acknowledgment

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Itinery

Mileage

0 Assembly point for trip is in the parking area on the southside of causeway Rt. 1, one mile east of Machias center at 8:00 a.m. Drive east on Rt. 1 to East Machias.

2.9 Turn right. Follow Rt. 191 to North Cutler.

12.4 Stop 1. Park along highway. This E-W trending end moraine is one of the most prominent and accessible in the area and is traceable as a nearly continuous ridge for at least 35 km (Fig. 1). In their study of the glacial geology of Maine, Leavitt and Perkins (1935) reported briefly on this region and on this particular end moraine that they named the Pond Ridge Moraine. Their description of the end moraine is as follows:

> "The frontal deposits (in Maine in general, and at Cutler in particular) take on the form of a rather smooth or hummocky ridge, with an ice-contact slope on the proximal side and a gently sloping wash plain on the distal side. Goldthwait has termed frontal deposits of this type moraine banks." Moraine banks were visualized as having formed along an ice margin which was standing in the sea.

The orientation of the shoreline exposure at this location provides an unusual opportunity to study the internal characteristics of the moraine. In general adequate exposures of the moraines of this region are notably absent.

The moraine rests (?) upon fossiliferous marine clay and is composed of interbedded sands and gravels, tills and fossiliferous marine, silty clay. Radiocarbon ages on the contained fossils date the formation of the moraine at approximately 13,500 years B.P. (Stuiver and Borns, unpub. data).

21.9 Return to Rt. 1 in East Machias via Rt. 191 and turn right. Pass through Machias center.

25.1 Stop 2. Turn right into Whitney's gravel pit.

This borrow pit has been excavated in a large moraine (Fig. 1) and exposes its internal character roughly from the moraine crest to its proximal side. The moraine is composed of stratified sand and gravel with interbeds of compact till. Deformation structures show translation from the proximal to the distal side of the moraine and probably reflect the thrusting of the ice sheet. No marine sediments have been recognized in this pit. However, at other locations marine silts are interfingered with the coarser sediments on the distal side of this moraine.

Return to Rt. 1 and turn right (west) and follow the crest of this moraine for approximately 3.5 miles.

34.0 Stop 3. Lunch on Carr Hill.

The several borrow pits on this hill expose stratified sand and gravel in long foreset beds dipping to the south. No till or marine silt are exposed. This hill, located along a moraine line (Fig. 1), is interpreted as an ice marginal kame probably formed in the sea. The top of the hill at altitude of 217 feet is below the upper marine limit of approximately 300 feet in the coastal zone. Continue west on Rt. 1.

48.2 Turn right on Rt. 191 in the center of Cherryfield (the Blueberry Capitol of the World).

49.5 Turn right on Ridge Road.
The road roughly follows the crest of a N.E.-S.W. trending end moraine whose proximal side faces northwest (to your left).

51.7 Stop 4. Park along the roadside. At this location the end moraine, along whose crest we've been driving, wraps around the NW side of the bedrock knob. From this location there is a panoramic view of a few of the hundreds of small, boulder strewn, recessional end moraines located to the northwest. These moraines are often clustered and may be called "washboard" moraines.

These moraines were formed along the margin of an ice lobe retreating to the northwest from a position southeast of Stop 4 (Fig. 1).

On the skyline to the north you can see the front slope of the Pineo Ridge ice-marginal delta extending for at least 6 miles east to west. The delta was prograded over the northern ends of many of the small moraines just discussed and therefore is younger than these moraines.

Continue northward for approximatly 3 miles on or parallel to this moraine and then drive up the partially disected front slope of Pineo Ridge delta. Note the fine sand in the road cuts. Low areas to the south of here are veneered with marine silt and clay of the Presumpscot Formation (Bloom 1960).

54.2 Top of delta. Bear left and drive north across the delta surface to the ice-contact proximal slope.

Note the myriad of abandoned distributary meltwater channels, the coarsening of the sediments toward the north, and the change from a smooth to a [kettled] surface as the proximal side is approached.

55.9 Stop 5. We are standing on the Pineo Ridge Moraine. This end moraine curves northwest from here where it ends (?) in approximately 10 miles. To the east it has been mapped, discontinuously

for approximately 60 miles and at Lubec, Maine passes into Passamaquoddy Bay. This moraine has not yet been recognized in New Brunswick. Return south on same road.

57.7 Turn left and drive east along the distal edge of the delta.

59.4 Stop 6. We are standing at the east end of the U.S. Coast and Geodetic Survey Base Line upon which most mapping control in Maine is based. This line was surveyed in the first half of the 19th century and Jefferson Davis, later to become president of the Confederate States of America, participated in the survey.

As you look around at this and other locations you will notice the vast ares of blueberry cultivation. In 1958 this area of Maine produced 85% of the world's commercial berries. Of the 100,000 acres of land in Maine presently under blueberry cultivation, 50,000 are in this immediate area. We can thank the Laurentide Ice Sheet for this!

This is the eastern end of the Pineo Ridge delta and we're standing near the outer edge of its depositional surface at an altitude of 260 feet. As the constructional phase of the delta came to a close the delta emerged about 20 feet above sea level and the sea eroded the prominent bench and cliff before us. This *nickpoint at an altitude of 240 feet is present along approximately 90% of the length of the delta front. Prominent nickpoints at this altitude have been recognized at various other locations in the coastal region from Lubec to Waldoboro, Maine, a distance of approximatley 180 miles. The local collapse of the 240 foot-high shoreline into kettle holes and the presence of meltwater stream channels graded to a sea level lower than the altitude of 240 feet on this delta indicate the presence of glacial ice in the area while emergence was in progress.

After this prominent nickpoint was formed the delta emerged relatively more rapidly and several lower, less conspicuous nickpoints were developed at lower altitudes. Relative sea level lowered to approximately -180 feet by 10,000 years B.P. and has subsequently risen to its present position.

* ("nickpoint" as used here refers to the point where the wave-cut cliff joins the wave-cut platform. It represents a former sea level.)

Drive east on the same road. Note that you are driving on the Pineo Ridge Moraine and that the delta is no longer present.

61.9 Turn right on paved road.

62.2 Stop 7. Park along the roadside.

At this location the E-W trending Pineo Ridge Moraine, constructed by ice readvancing from just east of north, crosscuts the N-S trending moraines constructed by ice that had advanced from and was receding toward the northwest.

Drive south to Rt. 1, turn left and continue to Ellsworth, Bangor and Orono. This will take approximately 1 1/2 hours.

SEDIMENTARY AND SLUMP STRUCTURES OF CENTRAL MAINE

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Introduction

The rocks of central Maine are turbidites and distal turbidites of predominantly Silurian age. They are metamorphosed to chlorite, or in places biotite, zone rocks. A variety of tectonic structures have been superimposed on the primary sedimentary and slump structures. The two processes of turbidite deposition and slumping tend to be associated because rapid deposition on a slope and water-saturation will aid both. Some structures observed are difficult to assign to either depositional or slump processes, and may be mistaken for some structures of tectonic origin (such as isoclinal folds and shears).

Acknowledgements

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Lithologic Units

Lithologies to be visited on Trip A-2 include the Vassalboro formation and an unnamed unit informally referred to as the "Kenduskeag formation". These units are described in detail by Ludman and Griffin in Trip B-4.

Distinguishing features of the Vassalboro formation are its massive and homogeneous character and thick to very thick bedding. Its composition is predominantly fine quartzite and coarse metasiltstone with minor phyllitic interbeds. Turbidite structures and slump structures are rarely observed in the Vassalboro formation, except at Peaked Mountain (Stop 1). Graded bedding and braided laminae are common.

Rocks of the "Kenduskeag formation" are characterized by extreme variation in bedding thickness and style, ranging from massive quartzites to zones of thin alternating metasiltstone and phyllite. Turbidite features are relatively common, as are slump folds and associated structures. An important characteristic of this unit is the chaotic zones in which bedding is totally disrupted.

Other units in central Maine are the "Sangerville formation", "Solon formation", and Waterville formation. The "Sangerville formation" exhibits excellent turbidite features and some slump structures. The "Solon formation" exhibits excellent turbidite features and locally well-developed sole markings. The Waterville formation exhibits fine internal laminae and cross-laminae, and good slump folds. The maroon and green phyllite member of the Waterville formation contains trace fossils, nerites facies (bathyal),similar to those described by Seilacher (1962, 1964, 1967).

Sedimentary Structures

In this paper, sedimentary structures are those structures associated with the deposition of the unit. They include features summarized in the first part of Table 1. Most of these features are the result of turbidite flow. Graded bedding and cross-laminae are common in rocks of central Maine, but the other features are relatively rare.

Slump Structures

Slump structures are those structures resulting from downslope movement prior to lithification (summarized in the second part of Table 1). The processes of slumping and of turbidite flow may produce similar structures in some cases.

<u>Slump Folds</u>

It is not always possible to distinguish isoclinal slump folds from tectonic isoclinal folds. Slump folds in central Maine usually occur as a set, that is, an anticline and a syncline, thus producing an asymmetric sigmoidal structure which appears as a minor structure superimposed on the major tectonic structures. Throughout the area slump folds are usually right-laterally asymmetric (z-folds, Ramsay, 1967, p. 351). Typically a slump fold set is not continuous through a thick stratigraphic sequence but tends to die out or to end abruptly. The stratigraphic thickness affected is generally 2.5 centimeters to 3 meters. A single outcrop 10 meters square in area may exhibit as many as eight different and distinct slump folds.

Slump folds appear to have deformed plastically while saturated with water. Individual beds often thicken in the crests and troughs, indicating that material flowed from one portion of the bed to another. The boundaries of most isoclinal slump folds consist of a sedimentary welded contact above and a sedimentary decollement surface below (Figure 1).

The criteria used to distinguish sedimentary slump folds from tectonic folds are: 1) that the slump fold is a flexural-flow fold formed by plastic deformation, 2) the shape of the fold changes from bed to bed, 3) the nose of the fold is usually blunt-shaped, but the folding is tight, 4) slump folds are usually found in a set of an anticline and a syncline, producing a disharmonic structure that is restricted to a stratigraphic horizon between undisturbed beds, 5) portions of a slump can be classed as a similar fold, 6) slump folds often have sedimentary welded upper contacts and sedimentary decollement lower contacts, 7) slump folds are usually associated with other soft-sediment flow structures, and 8) the axial lines of slump folds have a more variable plunge than those of tectonic folds.

<u>Sedimentary Welded Contacts</u>

Helwig (1970) proposed the term "welded contact", following the usage of Jones (1937), to describe a primary sedimentary relationship consisting of a penecontemporaneous depositional contact of continuous strata overlying a slump fold (Figure 1). The sedimentary welded contact is not a sharp, knife-edge contact; rather, the contact is indistinct, and internal

Table 1. Sedimentary, slump, and tectonic structures of central Maine.

Sedimentary Structures (resulting from, or associated with, turbidite flow)
- Graded beds
- Cross-laminae and cross-bedding
- Discontinuous bedding
  - Wedge-outs
  - Pull-aparts
- Sedimentary lenses
- Slate chips
- Siltstone balls and deformed pebbles
- Flow casts
- Sole markings

Slump Structures (resulting from submarine slumping and sliding)
- Slump folds
- Sedimentary welded contacts
- Sedimentary decollements
- Discontinuous bedding
  - Wedge-outs
  - Pull-aparts
- Sedimentary lenses
- Siltstone balls and deformed pebbles
- Chaotic bedding or sedimentary breccia

Dewatering structures
- Flame structures
- Load casts
- Braided laminae or "false bedding"

Tectonic Structures
- Tight, upright isoclinal folds (F_1) & axial plane cleavage (S_1)
- Open, asymmetrical folds (F_2) & both horizontal and vertical fracture cleavage (S_2)
- Open, small amplitude crenulations (F_3) (In Harmony area)
- Left- and right-lateral kink bands (F_4)
- Minor tectonic shears and faults (S_5)

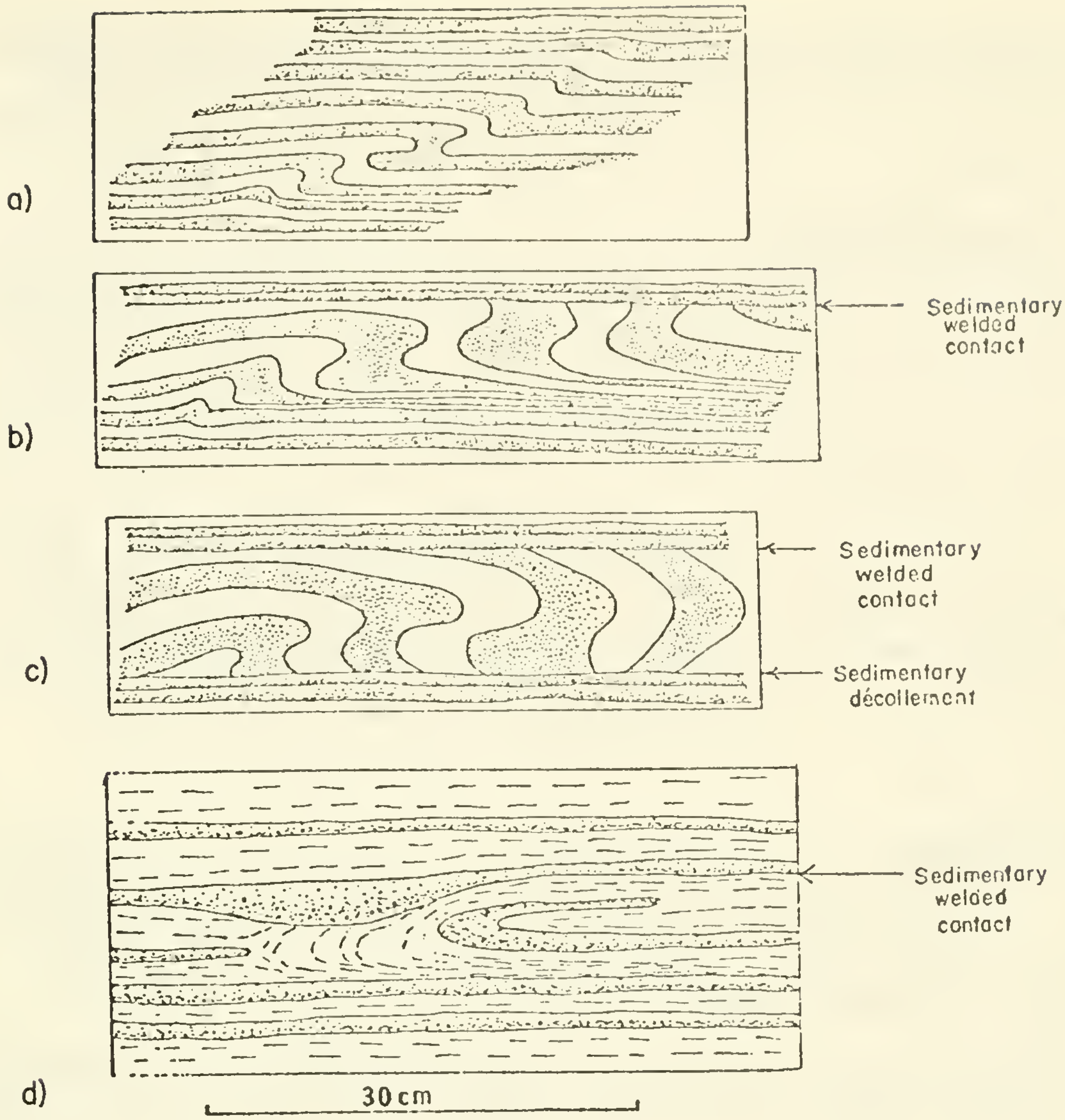

Figure 1. Features of slump folds.

(a) Slump fold which continuously changes its shape and dies out both upward and downward so that underlying and overlying beds are undeformed.

(b) Slump fold with a sedimentary welded contact at its top. The structure dies out to the lower left.

(c) Slump fold with a sedimentary welded contact at its top and a pronounced sedimentary décollement at its base.

(d) Diagram of flap fold in thin 0.3-1.2 cm interbedded phyllite (dashes) and quartzite (stipples) layers of the Waterville formation. Flap fold is overlain by a sedimentary welded contact. Quartzite layers are graded.

laminae are wispy and feathery where they intersect the contact, due to the unconsolidated nature of the sediments when the contact was formed. In most cases, a welded contact forms a zone 3 to 6 mm thick. Unlike tectonic shears and bedding plane faults, which exhibit a similar geometry, welded contacts do not weather more readily than the surrounding bedding planes.

Sedimentary Decollements

The lower contact of most slump folds is a surface of decollement along which the slump moved downslope (Figure 1c). This surface of sedimentary decollement is not a sharp distinct plane. Typically, laminae on either side of the decollement surface are feathery and wispy, indicating that the sediments were saturated and flowed plastically at the time of slumping. The lower limb of the slump fold has been overturned so that beds on either side of the decollement surface have opposing directions (Figure 1c).

In some cases the lower or overturned limb has been sheared completely off so that there is no reversal of top direction across the plane of decollement. In extreme cases, only the curved nose portion of the lower limb remains, exhibiting an angular relationship with the underlying beds. As in the case of a welded contact, the surface of sedimentary decollement does not produce an easily weathered zone across the outcrop, and thus is readily distinguished from a tectonic fracture zone.

Chaotic Bedding

Outcrops containing numerous deformed pebbles exhibit a chaotic aspect. In such outcrops, long axes of the pebbles are randomly oriented to subparallel. There may be no discernible bedding, the outcrop consisting entirely of deformed siltstone or quartzite pebbles, lenses, and segments of beds in a matrix of phyllite.

In some cases, the long axes of pebbles are parallel to the trend of bedding; these are often isolated lenses (sedimentary lenses of Table 1). Individual fragments within a chaotic zone range from 2.5 to 60 cm in length and the same chaotic zone may contain spherical pebbles, elongate pebbles folded back upon themselves, and flat tabular segments of beds. In many cases both the pebbles and the surrounding phyllite and fine siltstone exhibit extremely complex soft-sediment deformation (as at Stop 9).

In large exposures it can be seen that chaotic bedding or sedimentary breccia have developed in particular stratigraphic horizons which are about one to 2.5 meters thick, and continuous bedding can usually be traced along both sides of the chaotic zones.

Dewatering Structures

The authors believe that in central Maine, folding (F_1) and the development of cleavage (S_1) began while the sediments were still saturated with water, as suggested by Maxwell (1962) for the Martinsburg formation of New Jersey and Pennsylvania. Evidence for this is the

digitations on fold noses, the plastic nature of deformation, and the features referred to as braided laminae or "false bedding" (Table 1).

Braided Laminae

Black clay-size muscovite and chlorite grains are concentrated along the foliation planes of the medium-grained massive quartzites, producing a faint lamination. These foliation planes are separated by 6 to 12 mm, and the indistinct, faintly colored laminations are about 3 mm thick. The fine-grained material making up these laminae weathers more readily than the surrounding quartzite, thus producing a grooved appearance on the surfaces of weathered pavement outcrops.

In most cases it is difficult to determine whether the faint laminae of phyllite found in massive quartzites of the area are primary sedimentary features or are related to the foliation, because of the general parallelism between bedding and cleavage. In a few cases, such as at Stop 2, it can be conclusively demonstrated that the faint laminae are related to foliation (S_1). In these few cases the foliation is a few degrees divergent from bedding and the quartzite contains phyllite beds 10 to 20 cm thick. The laminae or "false bedding" intersect the trace of bedding at an angle of as much as 15^o.

Tectonic Structures

Five tectonic deformations have been recognized in central Maine. The first deformation (Acadian) produced most of the tight vertical isoclinal folds (F_1) seen in outcrop, and rotated the slump folds to vertical. It also produced the northeast-trending regional folds, and the vertical axial plane cleavage (S_1).

The second deformation produced the open right-laterally asymmetric folds (F_2) whose average orientation is N 15 E, vertical, and a complimentary horizontal set that is less well-developed. Associated with these folds are horizontal and vertical fracture cleavages (S_2), that often produce horizontal and vertical lineations (L_2).

The third deformation (F_3) is recognized only in the Harmony area of Skowhegan quadrangle (to be visited on Trip B-4, Ludman and Griffin).

The fourth deformation produced both left- and right-lateral kink bands (F_4), the majority being left-lateral. Their axial planes (S_4) are commonly oriented in an east-west direction.

The final stage of deformation is represented by minor but widespread tectonic shears (S_5). Probably the large-scale faults observed in the northern part of the area are related to these.

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Itinerary

Mileage	
	Meet in parking lot at Bangor Auditorium (behind the Paul Bunyan statue on Maine Street). Begin mileage at south exit of parking lot. Starting time is 8:30 A.M. The route as well as the regional geology are shown on Figure 2.
0	Turn east out of parking lot onto Dutton Street.
0.1	Turn south onto Main Street (Route 1A).
0.2	Turn west onto I-395.
0.7	Take the Hampden exit onto U.S. 202 West, and drive west.
3.2-3.5	Outcrops on west side of road of the "Kenduskeag formation".
4.8	Soudabscook Stream.
5.6	At stop sign, turn west onto Route 202/9 towards Unity.
7.4-7.7	Outcrops of Vassalboro formation on north side of road.
9.1	Soudabscook Stream.
9.3-13.3	Outcrops of massive Vassalboro formation.
14.8	View of Peaked Mountain straight ahead.
15.5-17.0	Outcrops of Vassalboro formation.
17.2	Dixmont town line.
17.7	Turn north onto dirt road at crest of hill and continue west on old asphalt road.
17.8	Turn north onto gravel road.
18.1	Proceed uphill to bend, turn around, and park on shoulder of road.

Stop 1. Peaked Mountain.
Pavement outcrops of Siluro-Devonian Vassalboro formation, consisting of low biotite zone massive quartzites with minor interbeds of blue-black phyllite and rusty-weathering blakc carbonaceous phyllite. A few brown-weathering calcareous lenses are present.

Sedimentary features include graded beds, flute casts (? or load casts, flame structures or injections, pelitic intraclasts, and cross-laminae.

Slump structures, near the south end of the outcrop,

include slump folds, wedge-outs, sedimentary decollements, and disrupted bedding. There is local development of "false bedding" or braided laminae parallel to axial plane cleavage (S_1).

Tectonic structures include an isoclinal syncline (F_1), two well-developed cleavages at about N 70 E (S_1) and N 10-15 E (S_2), a poorly developed cleavage of unknown origin at about N 35 E, and left-lateral kink bands (F_4).

Return to U.S. 202.

18.5	Turn west onto U.S. 202.
18.6-19.2	Outcrops of Vassalboro formation along the road.
19.8-19.9	Outcrops of "Kenduskeag formation" along the road.
20.8	View of Mt. Harris straight ahead.
21.3	Outcrops of Vassalboro formation.
21.7	Dixmont Hills picnic area.
22.0	Outcrop of Vassalboro formation south of road.
22.8	Town of Dixmont. Turn north on U.S. 7 towards Newport.
23.4-25.1	Outcrops of Vassalboro formation.
26.7	Plymouth town line.
27.9	Road crosses a kame, currently being mined for gravel.
28.1	Here the road follows an esker.
28.4	Plymouth Pond visable east of road.
29.6	Town of Plymouth. Turn sharp left and cross bridge above dam. Proceed southwest.
29.7	Pavement outcrop of "Kenduskeag formation" in house yard.
30.0	Turn west at "Y" (<u>away</u> from Round Pond) and proceed uphill.
31.2	Turn south onto dirt road.
31.5	Farm house. Park off road behind garage.

<u>Stop 2</u>. Plymouth Hill road traverse.
After viewing the outcrop behind the garage walk south along the woods road to the brow of the hill (about .4 miles), observing outcrops in the road and along both sides of it.

Pavement outcrops of Siluro-Devonian Vassalboro and Silurian "Kenduskeag formation". The latter consists of chlorite zone interbedded quartzites and blue-gray phyllites, characterized by zones of disrupted bedding and slump folds.

The first outcrop (Vassalboro formation) exhibits "false bedding" or braided laminae produced by injection of pelite along surfaces of axial plane cleavage at an angle to true bedding. Graded beds and left-lateral kink folds are also present.

The rest of the outcrops ("Kenduskeag formation") exhibit graded bedding, folded pebbles, slump folds, sedimentary breccia, pull-aparts, and wedge-outs. Note that the axial plane cleavage (S_1) cuts across the folded pebbles in the sedimentary breccia.

Other tectonic structures include folds (F_2), fracture cleavage (S_2), kink bands (F_4), and minor tectonic shears (S_5).

Although isoclinal folds are present it is not always possible to distinguish those of tectonic origin (F_1) from those of slump origin.

Return to asphalt road.

31.8 Turn east onto asphalt road and proceed slowly.

32.1 Just west of the Southern Gospel Mission Headquarters building pull as far off the road as possible and park.

Stop 3. Southern Gospel Mission outcrop.
A small pavement outcrop of "Kenduskeag formation".

Sedimentary structures include graded beds and cross-laminae. Slump structures include slump folds, sedimentary breccia, rotated pebbles, and sedimentary decollements.

Axial plane cleavage (S_1) is poorly developed, however, slump structures are deformed by open right-handed asymmetric tectonic folds (F_2).

Continue east on asphalt road.

32.7 View of Plymouth Pond and the Dixmont Hills.

32.9 At the "Y" bear left (north) towards Plymouth.

33.3 At stop sign in Plymouth bear east onto Route 169 towards Carmel.

33.3-35.0 Outcrops of "Kenduskeag formation".

35.1 Etna town line.

35.6 Outcrop of "Kenduskeag formation".

35.8 Dirt road to north. Continue east on Route 169.

35.9-37.8 Outcrops of "Kenduskeag formation". Open asymmetrical folds (F_2) visable in places.

38.0 Park on shoulder of road near large outcrop.

Stop 4. Etna interchange.
Roadcut outcrop of "Kenduskeag formation", with polished pavement outcrops on top.

Sedimentary features include graded beds and cross-laminae. Slump structures include slump folds, sedimentary breccia, balls and pebbles, pelitic injections, and pull-aparts.

Tectonic structures include axial plane cleavage (S_1), open asymmetric folds (F_2), both horizontal and vertical fracture cleavage (S_2), and tectonic thinning of beds (probably associated with F_1).

It is not clear in all cases whether the isoclinal folds and the breccia are of sedimentary or tectonic origin.

Continue east on Route 169.

38.8 Turn north onto State Highway 143.

39.0 Cross I-95 overpass.

40.3 At power line crossing park on shoulder of road, and walk back to the outcrop.

Stop 5. Power line outcrop.
Roadcut and pavement outcrops of a coarser facies of the "Kenduskeag formation".

This outcrop is distinguished for its excellent development of sedimentary breccia. The wispy nature of the pelite concentrations suggests deformation during dewatering similar to that discussed by Alterman (1973).

Calcareous pebbles with internal laminae can be observed on both glacially polished and blasted surfaces.

Continue north on State 143.

41.2 Intersection of State 143 and U.S. 2. Gas and food available. Turn east onto U.S. 2.

41.7 Outcrop of "Kenduskeag formation".

41.8 Carmel town line.

42.1 Turn right into Etna Pond Picnic Area.

Stop 6. Lunch.
The outcrops here are poor pavement exposures of the "Kenduskeag formation", typical of the bulk of outcrops in central Maine. Features which may be observed here are graded bedding and cross-laminae.

Turn west on U.S. 2.

43.0 Intersection with State 143; remain on U.S. 2.

43.7 Outcrop of "Kenduskeag formation".

43.9 Outcrop of "Kenduskeag formation" with isoclinal fold visible at southwest end of outcrop.

44.9 Outcrop of "Kenduskeag formation".

45.2 Just beyond crest of hill, bear left (southwest).

45.3 Turn southeast onto dirt road. Park on shoulder of road in vicinity of fourth telephone pole. The outcrops to be viewed are on the west side of the road between the fourth telephone pole and the top of the hill.

Stop 7. Washington School traverse.
Pavement outcrops of "Kenduskeag formation".

Sedimentary features here include graded bedding, cross-laminae, and possible flute casts or load casts.

Slump structures include discontinuous beds and lenses, sedimentary breccia, pelitic injections, slump folds, sedimentary decollements, and detached slump fold noses.

Tectonic features include axial plane cleavage (S_1), open asymmetrical folds (F_2), and kink bands (F_4).

It is uncertain whether many of the folds observed here are of slump or tectonic origin.

Return to U.S. 2.

46.0 Turn west onto U.S. 2.

46.7-46.8 Outcrops of Vassalboro formation.

47.4 Turn north onto side road at sign advertising the Sebasticook Valley Snowmobile Club.

47.5 Newport town line; outcrops of Vassalboro formation.

48.1 Maine Central Railroad crossing.

49.0 At stop sign turn northeast.

49.5 Outcrops of Vassalboro formation.

50.2 Stetson town line.

50.6 Stop by the small swamp (south of road) and park on shoulder. The outcrop is on the other side of the swamp.

Stop 8. Quarry outcrop.
Pavement outcrop of "Kenduskeag formation".

Sedimentary features include very good graded bedding, cross-laminae, convolute laminae, and possible load or flute casts.

Slump structures include slump folds, pull-aparts, wedge-outs, sedimentary decollements.

Tectonic features are poorly developed, except for the axial plane cleavage (S_1).

Return to vehicles and continue driving northeast.

52.5 Stop sign at intersection with State 143. Continue straight across.

53.6 Bear north (left) at "Y".

55.0 View of Pleasant Lake to left rear.

56.8 At stop sign turn east onto State Highway 222.

57.4 Levant town line.

59.6 Stop sign in town of West Levant. Leave State 222 and continue straight on Kenduskeag Road.

61.7 Kenduskeag town line.

63.9 Cross bridge over Kenduskeag Stream.

64.0 Turn south onto State Highway 15 at town of Kenduskeag. Gas available.

66.3 Outcrop of "Kenduskeag formation".

66.8 Turn east onto side road. Kenduskeag Stream visible to the south.

68.4 Bangor and Aroostook Railroad tracks.

68.9 Stop sign at intersection with Maine State Highway 221. Continue straight across.

70.7 Stop sign. Turn south.

71.3 View of Pushaw Lake to east.

72.3 Outcrop of "Kenduskeag formation". Turn east.

73.8 Stop sign. Go straight across intersection (towards Villa-Vaughn Campground).

78.2 Stop sign. Turn northeast onto Stillwater Ave.

79.0 I-95 overpass. Gas available here. Continue on Stillwater Ave.

79.5 Turn north onto State Highway 16 West (Bennoch Rd.) at blinking yellow traffic light.

80.2 Road is on esker.

81.9 View of Stillwater River and Marsh Island to east.

82.4 Esker.

83.3 Pushaw Stream.

85.2 Town of Pea Cove. Turn east onto Route 116 and proceed slowly.

85.4 Park on shoulder of road. Please do not park across driveways or on lawns. The outcrops to be viewed are in the field behind the row of houses.

Stop 9. Pea Cove.
Excellent pavement outcrops of "Kenduskeag formation".

Sedimentary features include graded bedding, cross-laminae, cross-bedding, and possible flame structures.

Slump structures include slump folds and dismembered slump folds, sedimentary breccia in all stages of formation, deformed and broken pebbles. Some clasts within the sedimentary breccia are themselves sedimentary breccia, suggesting that brecciation is a repetitive process. Pelitic injections in a variety of orientations, not parallel with tectonic structures, may also be seen.

Tectonic structures include isoclinal folding (F_1) shown by reversal of tops, open asymmetrical folding (F_2), axial plane and fracture cleavages (S_1, S_2), and tectonic shears (S_5) distinguished from sedimentary features by lack of pelite injections.

85.6 Return to State 16 and turn north towards Milo.

86.4 I-95 overpass. View of Alton Bog to the northeast.

86.5 Turn left onto I-95 South. All outcrops between here and Bangor are "Kenduskeag formation".

92.8	Stillwater Road exit. Orono and the University of Maine.
103.5	Exit onto I-395.
105.5	End of freeway. Turn immediately south onto Dutton Street and return to parking lot of Bangor Auditorium.

END OF TRIP

Thank you.

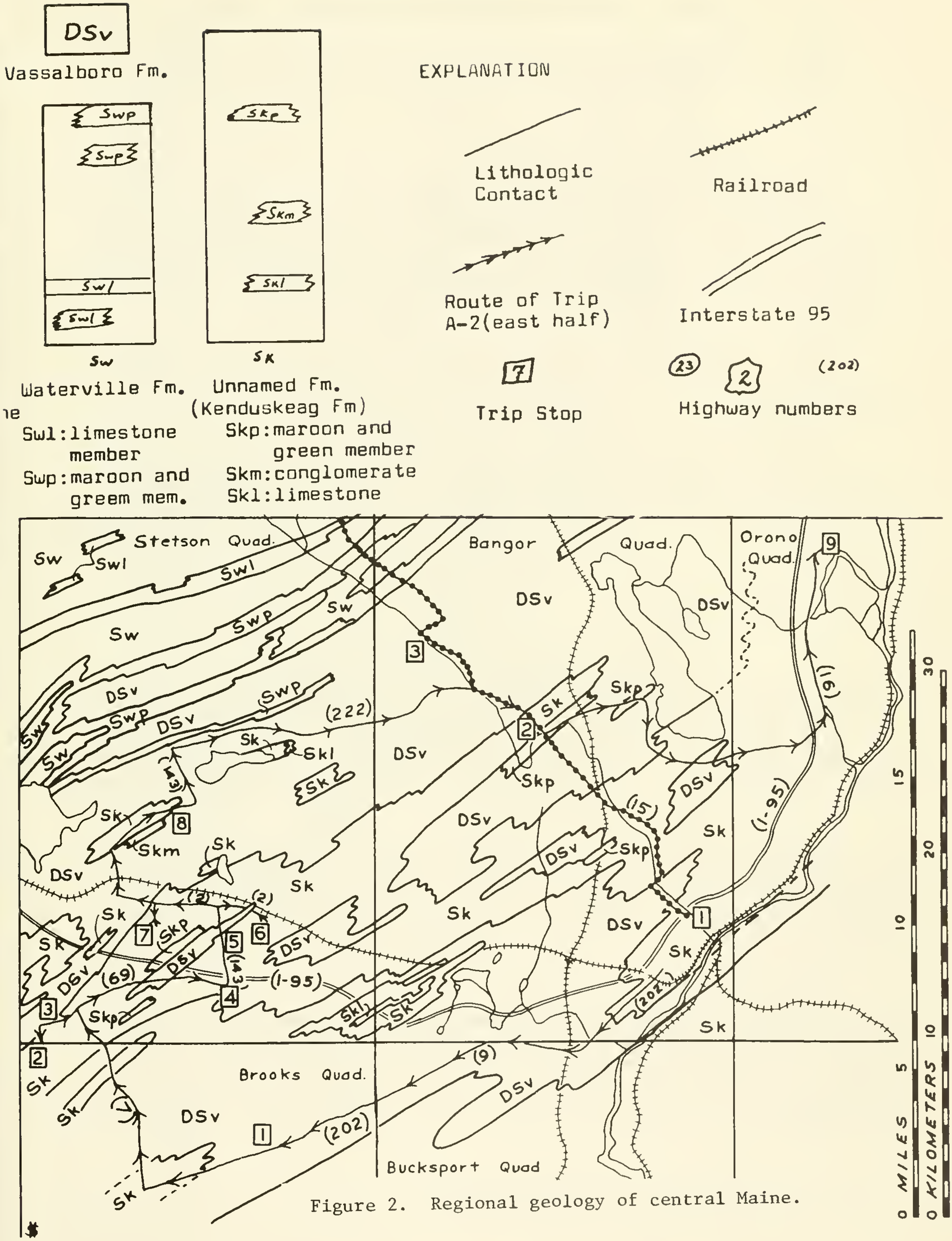

Figure 2. Regional geology of central Maine.

THE GEOLOGY OF THE CAMDEN-ROCKLAND AREA

Philip H. Osberg and Charles V. Guidotti†
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Introduction

The Camden-Rockland area encompasses the Camden, Rockland, Thomaston, and West Rockport 7 1/2 minute quadrangles of the U. S. Geological Survey Atlas of Topographic Maps. Geological mapping in the Camden (Osberg), the Rockland (Guidotti), and the Thomaston (Guidotti) quadrangles is essentially complete, but work in the West Rockport quadrangle (Osberg) is still in progress.

Rocks of the Camden-Rockland area are in the east flank of the Orrington-Liberty anticlinorium and belong to an uplifted block bounded east and west by longitudinal faults (see Foreword, this volume).

A simplified geologic map of the Camden-Rockland area is presented in Figure 1. Three fault-bounded blocks are recognized, each with a separate stratigraphy. The relationships within each block are complicated by a complex structural and metamorphic history. A large granite pluton intrudes the stratified rocks at the south boundary of the map-area and smaller bodies of granite and pegmatite are common in high-grade metamorphic terrains.

Stratigraphy

Three stratigraphic sequences have been worked out within the Camden-Rockland area. These sequences are identified as the Benner Hill sequence, the Rockport sequence, and the Megunticook sequence. Within each sequence the stratigraphic order has been established on the basis of sparse graded beds. The stratigraphic sequences and lithologic descriptions are tabulated in Figure 1.

The ages of these rocks in the Camden-Rockland area are thought to be Lower Paleozoic or older. Unit 2 of the Benner Hill sequence contains highly deformed brachiopods that have been assigned to the Ordovician (Boucot et al., 1972). Thus, the upper part of the Benner Hill sequence is Lower Paleozoic. The lower stratigraphic units may be Lower Paleozoic or older.

Unit 3 of the Megunticook sequence is lithologically similar to Middle Ordovician rocks exposed at Danforth (Larrabee, 1963) or to Lower Ordovician rocks near St. Stephens, N.B. (Neuman, oral communication, 1969). If either of these correlations are correct, unit 3 is Ordovician. Because of the absence of stratigraphic breaks in the Megunticook sequence, units 1 and 2 are thought to be either lowest Ordovician or possibly Cambrian, but not Precambrian.

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† Work supported by Maine Geological Survey

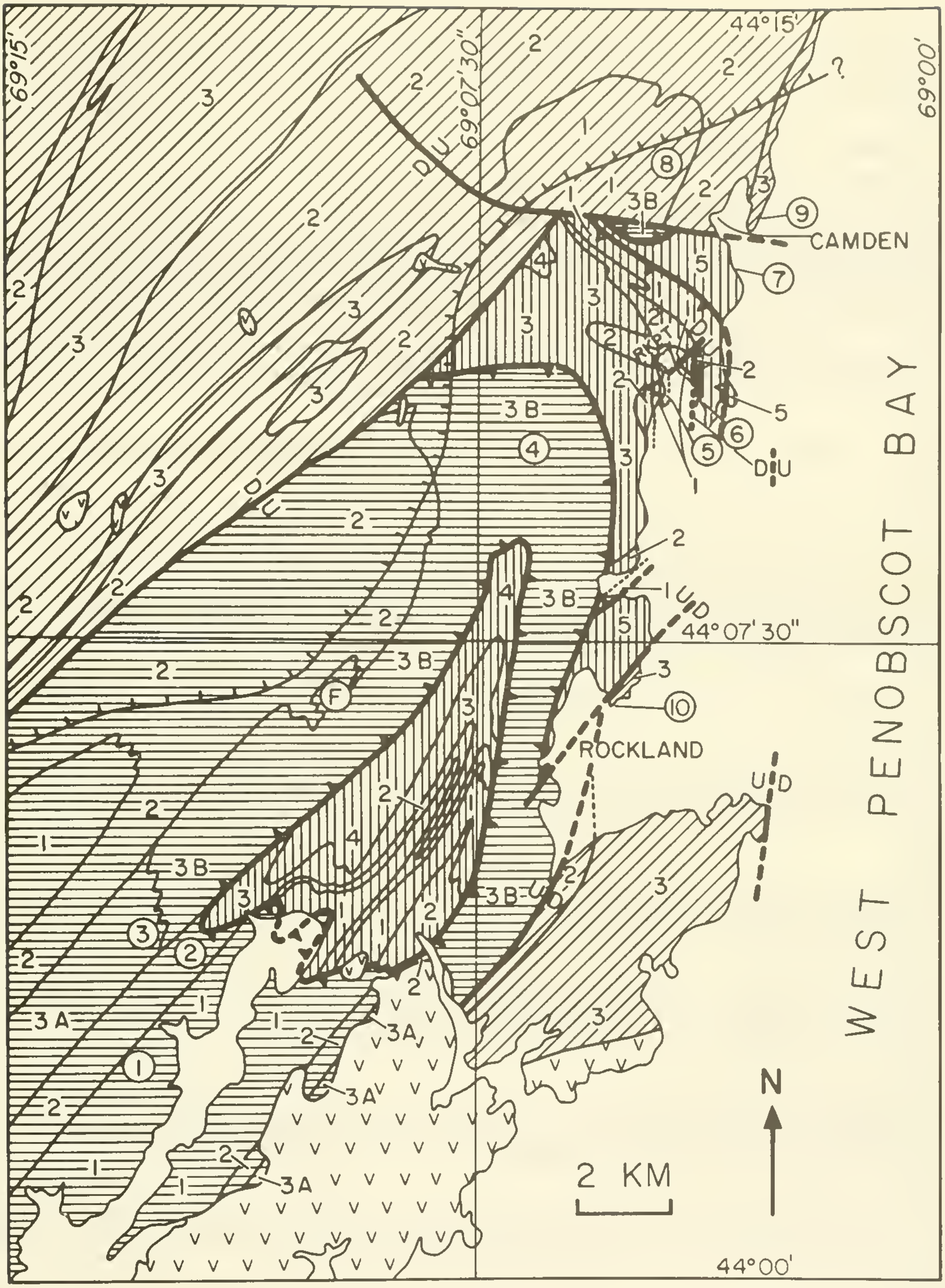

Figure 1. Generalized geologic map of the West Rockport, Camden, Thomaston, and Rockland 7 1/2' quadrangles.

EXPLANATION

INTRUSIVE ROCK

GRANITE AND PEGMATITE

BENNER HILL SEQUENCE

3A, 3B — 3A: RUSTY TO DARK GRAY QUARTZ-MICA SCHIST AND THIN QUARTZITE. LIGHT GRAY QUARTZ-MICA SCHIST AND FELDSPATHIC QUARTZITE. 3B: GRAY QUARTZ-BIOTITE SCHIST AND THIN GARNETIFEROUS QUARTZITE. SOME SANDY BEDS AND SPARSE CONGLOMERATE.

2 — LIGHT GRAY SCHISTOSE BIOTITIC QUARTZITE AND QUARTZ FELDSPAR-MICA GRIT. CONTAINS A RUSTY QUARTZITE AT UPPER CONTACT THAT CONTAINS DEFORMED BRACHIOPODS.

1 — THIN BEDDED, LIGHT COLORED FELDSPATHIC QUARTZITE AND GRAY QUARTZ-MICA SCHIST.

ROCKPORT SEQUENCE

5 — WHITE MARBLE. LIGHT GRAY QUARTZ-MICA-GARNET-ANDALUSITE SCHIST CONTAINING LOCALLY COARSELY GRADED BEDS OF GRANULE CONGLOMERATE AND LARGE LENSES OF POLYMICTIC CONGLOMERATE.

4 — WHITE MARBLE OCCURRING IN BEDS 4-12 CM. THICK, LOCALLY CONTAINING FINE GRAINED TREMOLITE.

3 — LIGHT GRAY QUARTZ-MICA-GARNET-ANDALUSITE SCHIST.

2 — WHITE, GLASSY QUARTZITE. QUARTZITE CONGLOMERATE. GRAY, BIOTITIC QUARTZITE AND CONGLOMERATE. DISCONFORMITY SEPARATES WHITE QUARTZITE AND GRAY BIOTITIC QUARTZITE.

1 — THINLY BEDDED BUFF MARBLE AND GRAY LIMESILICATE QUARTZITE. BUFF LIMESTONE CONGLOMERATE. GRAY QUARTZ-MICA SCHIST.

MEGUNTICOOK SEQUENCE

3 — RUSTY, DARK GRAY QUARTZ-MICA SCHIST AND BIOTITIC QUARTZITE. MINOR WHITE MARBLE AND GREENSTONE / AMPHIBOLITE.

2 — LIGHT GRAY QUARTZ-MICA-GARNET-ANDALUSITE / SILLIMANITE SCHIST. MINOR RUSTY. QUARTZ-MICA SCHIST AND QUARTZITE.

1 — WHITE TO LIGHT GRAY QUARTZITE AND QUARTZITE CONGLOMERATE.

FORMATION CONTACT

THRUST FAULT

U D HIGH ANGLE FAULT

LINE SEPARATING PROGRADE AND RETROGRADE METAMORPHISM

F FOSSIL LOCALITY

3 SCHEDULED STOPS

RKPT = TOWN OF ROCKPORT

The age of the Rockport sequence is more uncertain. It is presumed to be Lower Paleozoic or older because it is deformed by the same structural features that deform the other two sequences but which are cut by plutons that are thought to be Middle Devonian on the basis of lithic similarity to dated plutons (Faul et al., 1963; Zartman et al., 1970).

Lithic correlations between the three sequences cannot be made within the Camden-Rockland area. Both the Benner Hill and the Megunticook sequences contain units that have been assigned to the Ordovician, but the lithic aspects of these two units are different. They may represent different sedimentary facies within the same depositional basin, or, alternatively, the Benner Hill and associated Rockport sequences may bear a "Made in Africa" label.

Structural Geology

High-angle faults. Seven high-angle faults are delineated in Figure 1. These faults belong to two groups according to their ages. The east-trending fault through Camden, the northeast-trending fault that extends to the west edge of Figure 1, and the northeast-trending fault south of Rockland post-date the major folding but pre-date the prograde metamorphism and the intrusion of the large body of granite (Middle Devonian?) at the south edge of the map.

The other high-angle faults are younger. They commonly have silicified zones and slickensides associated with them. They are later than the major folding and the prograde metamorphism, and the fault at Owls Head (east of Rockland) cuts granitic intrusions (Middle Devonian?) and mafic dikes (Triassic?) as well.

The displacement on the older faults is such as to juxtapose the Benner Hill and Rockport sequences against the Megunticook sequence. Stratigraphic determinations within the Megunticook sequence constrain the displacement on the east-trending fault through Camden as is shown in Figure 1. The displacements on the other two early faults are not constrained locally. However, because of the possible continuity in stratigraphic section between the Megunticook sequence and the Silurian section of east-central Maine (see Foreword, this volume) and because of the wide geographic distribution of these sequences, the Benner Hill and Rockport sequences are thought to lie at a deeper tectonic level and are exposed as horsts. The displacements are so indicated on Figure 1. This model requires that a considerable component of right-lateral displacement exists on the east-trending fault through Camden and that the horst continues in a northeast direction toward Isleboro north of this fault (see Foreword and Trip A-6).

The displacements on the younger faults are constrained by the local stratigraphy.

Thrust faults. Two thrust faults have been mapped in the Camden-Rockland

area (Figure 1). One separates the Benner Hill sequence from the Rockport sequence, and the other divides the Rockport sequence into two parts. The thrust beneath the Benner Hill sequence is necessitated by the truncation of lithic units in both the Benner Hill and Rockport sequences along the trace of the fault. The thrust surface in a general way dips toward the west or south, but in detail is folded with northeast trends. An antiform produces a window into the Rockport sequence west of Rockland and a parallel, tight synform lies east of the window. This thrust surface is above the present erosion surface west of Rockport but again is truncated by the present surface of erosion to the north against the east-trending high-angle fault that passes through Camden.

The second thrust lies entirely within the Rockport sequence and is truncated by the thrust previously described. It places into contact unit 1 and unit 5, separating two sequences with opposite facing directions. Locally, its trace is marked by a pre-schistosity breccia. It is interpreted as a slide along the overturned limb of a recumbent anticline. The average dip of this thrust is thought to be toward the west or south, at a moderately steep angle although in detail it must be considerably contorted by later folding.

The thrust that lies entirely within the Rockport sequence is thought to be considerably older than that underlying the Benner Hill sequence. The thrust underlying the Benner Hill sequence truncates the thrust within the Rockport sequence, and it is later than the shistosity whereas the thrust within the Rockport sequence has an associated breccia that predates the schistosity.

Minor folds. Minor folds of the Camden-Rockland area have been assigned on the basis of interference characteristics and orientation to two eipsodes of deformation. The earliest folds deform only bedding and are moderately open to isoclinal and overturned to recumbent in style. Schistosity or segregation banding is parallel to their axial surfaces. Their plunges have a range of orientations.

Younger folds deform bedding, schistosity, segregation banding, and early folds. These folds have styles that are generally open, asymmetric, and overturned. A well developed cleavage parallels their axial surfaces and their plunges show a wide variation. Commonly their axial surfaces have fairly constant orientations, but analysis of them indicates on a regional scale that they are folded by still younger folds.

Although a third and younger set of folds can be shown to exist regionally, few minor folds can be related to them unambiguously.

Major folds. Major folds belong to three episodes of deformation - an early reclined folding, a folding of intermediate age with steeply dipping axial surfaces, and a still younger folding again with steeply dipping axial surfaces. Over much of the area the two younger sets of major folds have nearly parallel trends and cannot be separately distinguished.

An early reclined fold is delineated by the detailed stratigraphy

of unit 1 of the Rockport sequence (Fig. 1). The nose of this fold lies against the thrust fault from east of and to the northwest of Rockport. Units 1, 2, 3, and 4 of the Rockport sequence lie in its normal limb, and its inverted limb has been cut out along the thrust. This fold is thought to extend beneath the thrust sheet carrying the Benner Hill sequence, and it is present in the window west of Rockland where its axial trace is defined by unit 1 of the Rockport sequence. Units 2, 3, and 4 west of the axial trace are in the normal limb. This is corroborated in one of the pits of the Dragon Cement Company where unit 4 can be observed to overlie units 2 and 3. Unit 2 southeast of the axial trace is interpreted to be in the inverted limb.

A second reclined fold exposes unit 2 of the Megunticook sequence in the northwest part of the Camden-Rockland area, but the detailed geometry and structural relations have not yet been worked out for it.

Folds of intermediate age deform the normal limb of the early reclined fold in the vicinity of Rockport (Fig. 1). These folds have axial surfaces that dip steeply east and northeast. Their west limbs are nearly vertical and their east limbs dip gently. Plunges are generally toward the northwest. Minor folds associated with these folds have a right-handed sense on both limbs and are regarded as relics of the earlier reclined fold.

Folds in the Benner Hill sequence developed earlier than the underlying thrust and, therefore, are earlier than the folds that deform the thrust surface and are regarded as of intermediate age. These folds are overturned toward the east and have axial surfaces that strike northeast. Their plunges are also toward the northeast. The folds in the vicinity of the fossil locality indicated on Figure 1 may have an intermediate age as well.

The folds in the Rockport sequence that are exposed in the window through the Benner Hill sequence deform the early reclined fold. Their trends and styles are similar to those assigned to the intermediate folds and their geometry indicates that they were present before being overridden by the thrust. Consequently, they are assigned an intermediate age.

Other northeast-trending folds have orientations and styles consistent with the intermediate folds. The outcrop of unit 1 of the Megunticook sequence west of Camden (Fig. 1) delineates a large northeast plunging anticline. Unit 3 and the detailed stratigraphy within unit 2 delineate the same anticline to the northeast. The "canoe-shaped" outcrop of unit 3 of the Megunticook sequence west of Rockport is also consistent with the trend and style of the intermediate aged folds. Both of these structures are overturned toward the east and have northeast axial trends.

The youngest folds are unambiguously identified only in the vicinity of Rockport and in the Benner Hill thrust plate. The thrust surface beneath the Benner Hill sequence is folded into northeast trending antiforms and synforms that post-date the folds in the Benner Hill sequence. These youngest folds also deform rocks of the Rockport sequence in the vicinity of Rockport causing deflections in the attitudes of the axial surfaces of the intermediate aged folds and the early thrust. These folds, in general, are

more open than folds of intermediate age. Elsewhere the main effect of the youngest folding was to flatten preexisting folds of intermediate age.

Metamorphism

The rocks of the Camden-Rockland area have been recrystallized in a complex pattern suggesting the possibility of two different episodes of regional metamorphism. Southeast of the hatchured line in Figure 1 the rocks have abundant retrograde features, whereas to the northwest of this line, the metamorphic features are consistent with prograde recrystallization. Within the area of retrogradation a relict garnet and relict andalusite zone are recognized. The original metamorphic gradients increased toward the west and north. Within the area of progradation, a sillimanite and a K-spar + sillimanite zone are recognized. The gradients in this area increase toward the south and west.

Plutonic Rocks

The plutonic rocks of the Camden-Rockland area are of two ages. A few dikes of biotite granite, pegmatite, and gabbro have been folded by intermediate aged folds and the gabbroic dikes at least have been metamorphosed. Presumably the dikes of biotite granite and pegmatite have also been metamorphosed but the metamorphic effects are not apparent.

The large pluton of binary granite at the south edge of the map area (Fig. 1) is younger than the folding, early high-angle faulting, and prograde metamorphism, but earlier than the retrograde metamorphism. Smaller stocks, and dikes of binary granite and most pegmatites also have similar structural relationships. On the basis of lithic similarity to radiometrically dated plutons (Faul et al., 1963; Zartman et al, 1970) the younger igneous rocks are thought to have a Middle Devonian age.

Hornblende granite (not separately shown on map) at the south boundary of Figure 1 is of equivocal age. It occurs as inclusions in the binary granite, and therefore must be older than the binary granite. What these relationships mean in terms of the relative ages of the two granites is unclear.

Certain gabbroic and diabasic dikes have chilled margins and are younger than the retrograde metamorphic event. These may be as young as Triassic.

Geologic History

The Camden-Rockland area has undergone a complex series of events. Their sequence, as currently understood, is tabulated below:

1. Deposition of the Benner Hill, Rockport, and Megunticook sequences.
2. Early recumbent folding and the concomitant development of thrusts.
3. Intrusion of granite, pegmatite, and gabbroic dikes.
4. Development of folds with northeast trends.
5. Thrusting.
6. Further folding with northeast trends and flattening of earlier northeast folds.
7. Development of early high-angle faults.
8. Prograde metamorphism.
9. Intrusion of binary granite and associated dikes and pegamatites.
10. Second metamorphic event.
11. Intrusion of mafic dikes.
12. High-angle faulting.

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Larrabee, D.M. and C.W. Spencer, 1963, Bedrock geology of the Danforth quadrangle, Maine: U.S. Geol. Survey GQ-221.

Zartman, R.E., P.M. Hurley, H.W. Krueger, and B.J. Giletti, 1970, A Permian disturbance of K-Ar radiometric ages in New England: its occurrence and cause: G.S.A. Bull. 81, 3359-3374.

Itinerary

Mileage	
0	Assembly point for trip is in the parking area at the Chamber-of-Commerce building, Rockland, Maine. Starting time 9:00 A.M. Drive west on Route 1 to Thomaston.
4.4	Turn left on Wadsworth Street.
5.3	Cross Saint George River.
8.0	<u>Stop 1</u>. Park along highway. Walk through yard and approximately 950 feet along farm road. Unit #1 of the Benner Hill Sequence. The exposure is light gray,

thinly layered quartz-feldspar granulite and quartz-feldspar-biotite schist. Dikes of early biotite granite cut the layered rocks and both layered rocks and dikes are deformed by intermediate folds. Small displacements and shears are common.

Return to cars promptly. Reverse direction and proceed north.

8.7 Turn left on side road.

9.0 Stop 2. Park along roadside. Outcrop is approximately 500 feet west in field.

Unit #2 of the Benner Hill Sequence. Small outcrop of light gray, thinly bedded quartz-feldspar-biotite granulite and quartz-mica schist. An isoclinal fold is interpreted to be an intermediate fold. A pervasive slip cleavage is parallel to the axial surface of the fold. The slip cleavage is difficult to distinguish from bedding except in the nose of the fold. Beds with sand-sized clasts may be observed in an adjacent outcrop. A north-trending late pegmatite dike cuts the structures of intermediate age.

Return to cars. Continue north on road.

9.7 Stop sign. Bear left across Route 97 onto poorly paved farm road.

10.1 Stop 3. Park in yard at farm house. Outcrop is approximately 270 feet west of house and north of road.

Unit #3A of Benner Hill sequence. Light gray to rusty thin bedded quartz-mica schist containing pseudomorphs of andalusite and incompletely altered staurolite. Interbeds are quartz-feldspar-biotite granulite, quartzite, and isolated, boudinage blocks of calcsilicate granulite. The calcsilicate granulite beds are zoned with selvages of actinolite and cores of quartz, plagioclase, and grossularite. The large synform displayed by bedding is thought to be an intermediate fold. A slip cleavage in the schist is approximately parallel to the axial surface of the fold. Some of the quartzose beds are displaced along shears that strike north and dip steeply.

Exposure to the east contains rusty to gray quartz-mica schist interbedded with thin quartzose beds. Pseudomorphs after andalusite are present in some of the schistose beds and garnet and biotite are present in some of the quartzose beds. Folds similar to those previously described are well displayed. In addition the bedding is complexly deformed by folds that predate those described above. These earlier folds may be "slump folds". The outcrop is cut by early greenstone dikes.

Return to cars and drive back to Route 97. At Route 97, turn left and drive north.

11.6 Turn right onto Route 1. Continue north on Route 1 through Thomaston.

14.8 Turn left onto Old County Road.

16.3 Quarries along road are in Unit 4 of the Rockport sequence.

18.8 Stop sign. Continue north on Old County Road across Maverick Street.

20.8 Stop sign. Rejoin Route 1 and drive north.

20.9 Turn left on South Street. Caution of traffic.

21.7 Bear right at fork in road.

22.3 Stop sign. Continue on South Street across Rockville Street.

22.8 Stop 4. Park along side of road.

Unit #3B of the Benner Hill sequence. Exposure shows medium gray quartz-biotite schist in beds 8" to 15" thick. Some beds contain pseudomorphs of andalusite and pits that suggest cordierite(?). Other beds have a sandy texture. Beds, 1/2" to 1" thick, of dark gray quartz-garnet granulite separate the schistose beds. Locally light colored amphibole also occurs in the quartzose beds. An open fold is delineated by the quartzose beds.

Ordovician brachopods have been found in a calcareous quartzite bed that occurs stratigraphically just below this unit in the northern part of the Thomaston quadrangle (see Fig. 1). Although these fossils have great potential for dating purposes, the separation by faults of the stratigraphic sequence that contains the fossils from other geologic sequences has thus far restricted their usefulness.

Return to cars and continue north on South Street.

23.6 Turn right on tar road.

25.0 Stop sign. Turn right onto Route 1.

25.1 Turn left onto Pascals Avenue.

25.2 Turn right onto School Street.

25.3 Turn left onto Spruce Street.

25.5 Turn left onto Sea Street.

25.6 Stop 5. Park behind white house and along street. Walk down gravel driveway and right onto grassy path. Outcrop is at end of path. Caution rocks are apt to be slippery!

Unit #2 of the Rockport sequence. White, glassy quartzite. Some beds contain quartzite clasts that are flattened in the surface of bedding. Small folds in bedding are intermediate structural features.

At cove to south, white, glassy quartzite and conglomerate overlie thinly bedded marble and calcsilicate quartzite of Unit #1. Minor folds at the contact have the wrong sense for the closure of the intermediate fold displayed in the cove. This exposure is interpreted to be in the normal limb of an intermediate aged anticline and the minor structures are relics of the early folding.

Return to cars. Reverse direction and proceed to north on Sea Street.

26.0 Turn right on Pascals Avenue. Continue through center of Rockport.

26.6 Bear right at triangle.

26.7 Turn right on Mechanic Street.

27.0 Continue onto dirt-road extension of Mechanic Street.

27.6 Stop 6. Park on dirt road. Walk to exposure on shore.

Unit #1 of the Rockport sequence. Eastern end of outcrop is buff weathered, light gray limestone-clast conglomerate with limestone matrix. Clasts are only slightly deformed. Western part of outcrop is interbedded marble and dark gray biotite quartzite containing limesilicate minerals. Folds with axial surfaces that vary in orientation from steep to nearly horizontal are thought to be early folds. The variation in orientation is due to folding about a younger fold axis (approximately 30° N.27°E.). Siliceous deposition along slip cleavage causes the slip cleavage to stand out in relief. The slip cleavage is parallel to the axial surfaces of intermediate folds. Small faults locally displace bedding.

Return to cars and continue east on Mechanic Street.

27.7 Turn left onto dirt road.

28.0 Turn left onto tarred road.

28.5 Stop sign. Turn right onto Chestnut Street.

29.2 Turn right onto Bayview Street.

29.8 Stop 7. Park along road. Outcrop is to north on shore. Proceed down driveway, west of house, and down steps. Please respect the property.

Unit #5 of the Rockport sequence. Exposure is light gray quartz-mica-garnet-andalusite schist. Much of the andalusite has been altered to muscovite. Andalusite is concentrated along gently northeast dipping surfaces that are interpreted as segregation banding formed along an older cleavage. Bedding is indicated by granule conglomerate grading to fine sand and pelite on the north side of outcrop. These grades suggest that stratigraphic tops are northeast. Lenses and stringers of garnet rich quartzite in the schist parallel the bedding. A late fold involving the granule conglomerate and sandstone can be seen at the top of the exposure.

Return to cars and continue along Bayview Street.

30.9 Stop sign in center of Camden. Turn right onto Route 1.

31.1 Turn left onto Route 52.

31.4 Turn right on Trim Street.

31.5 Turn left on Megunticook Street and proceed to end of street.

31.5 Stop 8. Park in parking space and walk up trail approximately 540 feet.

Unit #1 of the Megunticook sequence. Ledges of massive, glassy, white to light gray conglomeratic quartzite (Battie Formation). Clasts are quartzite or biotitic quartzite and are subrounded. They are little deformed except for some fracturing. The matrix is biotitic quartzite.

Return to cars. Reverse direction and drive east on Megunticook Street.

31.6 Turn right.

31.7 Turn left on Route 52.

32.0 Turn left on Route 1.

33.0 Turn right on Sherman Point Road.

33.7 Bear right.

33.8 Stop 9. Turn cars around and park along road. Follow driveway to shore and walk along shore to Northeast Point.

Exposures along shore are of units #2 and #3 of the Megunticook sequence. The contact between the two units is in the small cove and can be seen at low tide.

Unit #2 consists of light gray quartz-mica-garnet-andalusite schist. Commonly the andalusite is wholy or partly pseudo-

morphed by muscovite. The garnet occurs distributed throughout the rock and also as coticules. Locally beds of biotitic quartzite, as much as a foot thick, are intercalated with the schist.

Unit #3 consists of rusty, medium quartz-mica-sulfide schist, some beds of which contain pseudomorphs of andalusite. A thin unit of white marble occurs just above the base of the unit #3, and higher in the section biotitic quartzite beds become prominent. The folds in the marble are interpreted to belong to the intermediate episode of folding.

Return to cars and retrace route along Sherman Point Road.

34.6 Turn left onto Route 1. Continue on Route 1 through Camden, to Glen Cove.

41.7 Turn left onto Warrenton Street in Glen Cove.

42.3 Turn right.

42.9 Turn left onto Samoset Street and proceed to end of street.

43.5 Stop 10. Park cars in parking area. Do not block roadway. Follow path to shore and walk west along shore approximately 2000' to the second point.

Unit #3 (?) of the Megunticook sequence (?). Exposure of light gray to gray biotitic quartzite in beds 1/2" to 4" thick. A few beds contain limesilicates. Intercalated are quartz-feldspar-mica granulites that may represent metavolcanics. Also present is a prominant unit of white to gray marble. Bedding is deformed by broad, open folds the orientations of which are consistent with the intermediate generation of folds. A prominant slip cleavage is essentially parallel to the axial surfaces of these folds. The marble unit is deformed by an early recumbent fold that is overturned toward the east. This early fold is deformed by folds and cut by slip cleavage of the later deformation. Numerous small faults off-set the bedding. A basaltic dike intrudes the layered rocks and cross-cuts the slip cleavage belonging to the late deformation. The dike, however, is off-set by the small faults, thus, dating the small faults as considerably younger than the late stage of folding. Outcrops along the shore to the east of this exposure are dominantly gray, quartz-feldspar-biotite, amphibole schist containing clots of chlorite or biotite. This rock may represent a meta-andeside in unit #3. Other rock-types include biotitic quartzite and rusty, gray quartz-mica schist.

Return to cars and proceed to Orono.

GENERAL BEDROCK GEOLOGY OF NORTHEASTERN MAINE

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Introduction

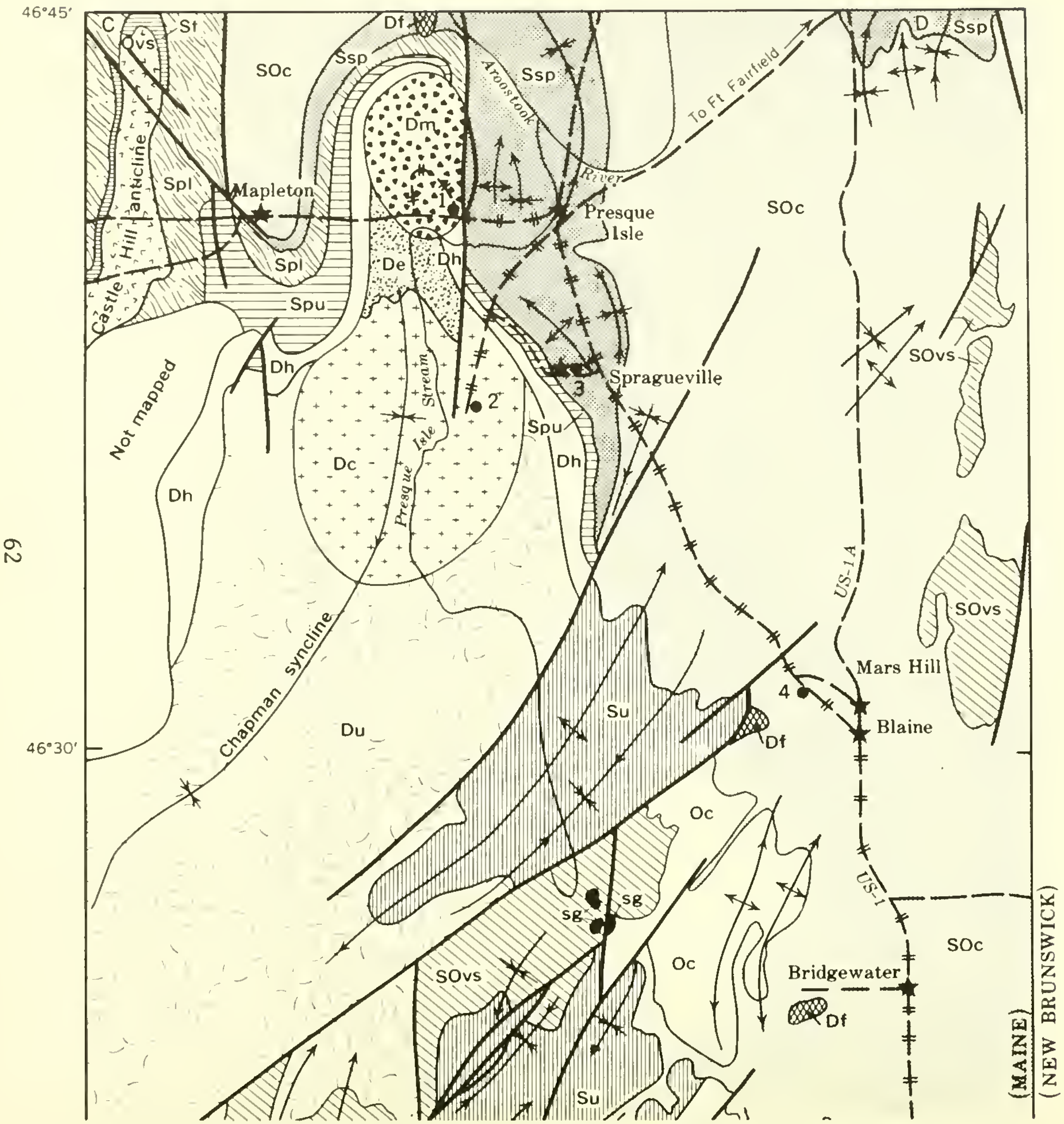
The purpose of this trip is to obtain an overview of the regional bedrock geology of northeastern Maine. Only readily accessible stops have been selected. These will show the characteristic lithologic features of formations that are important in defining unconformities or the absence of such unconformities, as well as the general tectonic features of the area. Much of the regional setting is described in the road log between stops; this part of the log should be closely followed. The geology of the region shown on figure 1, only a part of which is covered by the trip itinerary, has been described in several publications, especially those by Boucot and others (1964); Pavlides (1965, 1968, 1971, 1972, 1973); Pavlides and Berry (1966); and Pavlides, Boucot, and Skidmore (1968).

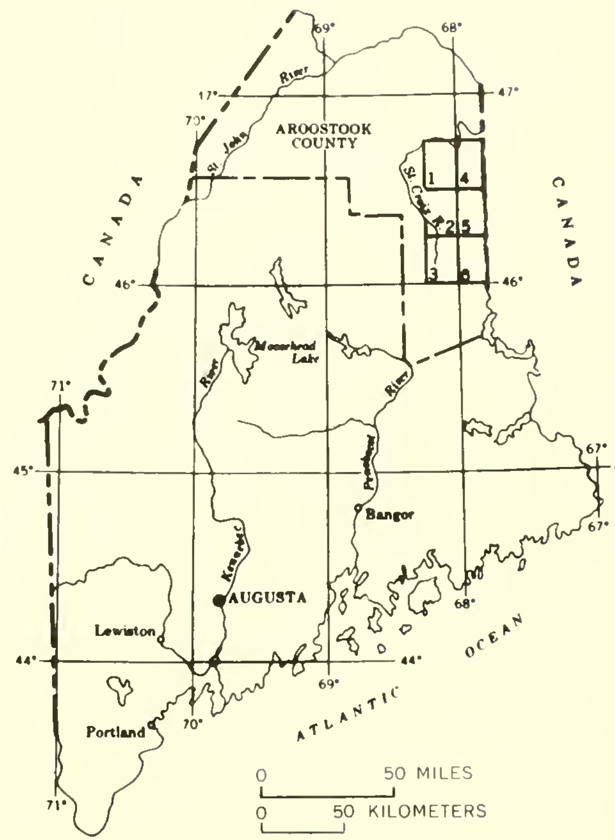
Stratigraphy and tectonics

Northeast Maine (fig. 1) is underlain by folded sedimentary and volcanic rocks that, in part, have a weak regional metamorphic imprint and that span Cambrian(?) through Middle Devonian time. Because the regional metamorphism is weak, sedimentary and metamorphic nomenclature are used almost interchangeably in parts of the report. Locally postkinematic, chiefly felsic plutons of Devonian age with well-developed thermal aureoles (Stops 10 and 11) intrude some of the folded rocks. Because of the presence of fossiliferous strata at various places, a relatively well-documented stratigraphy has been established locally so that paleontologic data support facies and regional tectonic interpretations to a greater degree than in most other areas in the New England Appalachians.

Stratigraphy.--The eastern part of the area (fig. 1) is underlain by limy rocks of the Carys Mills Formation of Middle Ordovician (Barnveldian) age (Graptolite Zone 13 of Berry, 1960, 1970) through Early Silurian (Llandovery) age (British Graptolite Zone 19). The Carys Mills contains turbidite features (Stops 4 and 5) and is considered to be a variety of calcareous flysch. The Carys Mills and its temporal equivalents north of the area shown in figure 1 form the core of the Aroostook-Matapedia anticlinorium that extends from about Oakfield, Maine, to near Percé at the east end of the southern part of the Gaspé Peninsula of Quebec. The limy rocks of the Carys Mills are the basis for the intensive agricultural development of Aroostook County.

These calcareous rocks were probably deposited in the central parts of a trough, herein designated as the Carys Mills basin, that is nearly coextensive





1, Presque Isle
2, Howe Brook
3, Smyrna Mills
4, Mars Hill
5, Bridgewater
6, Houlton

Index showing location of map area and quadrangles

A-5

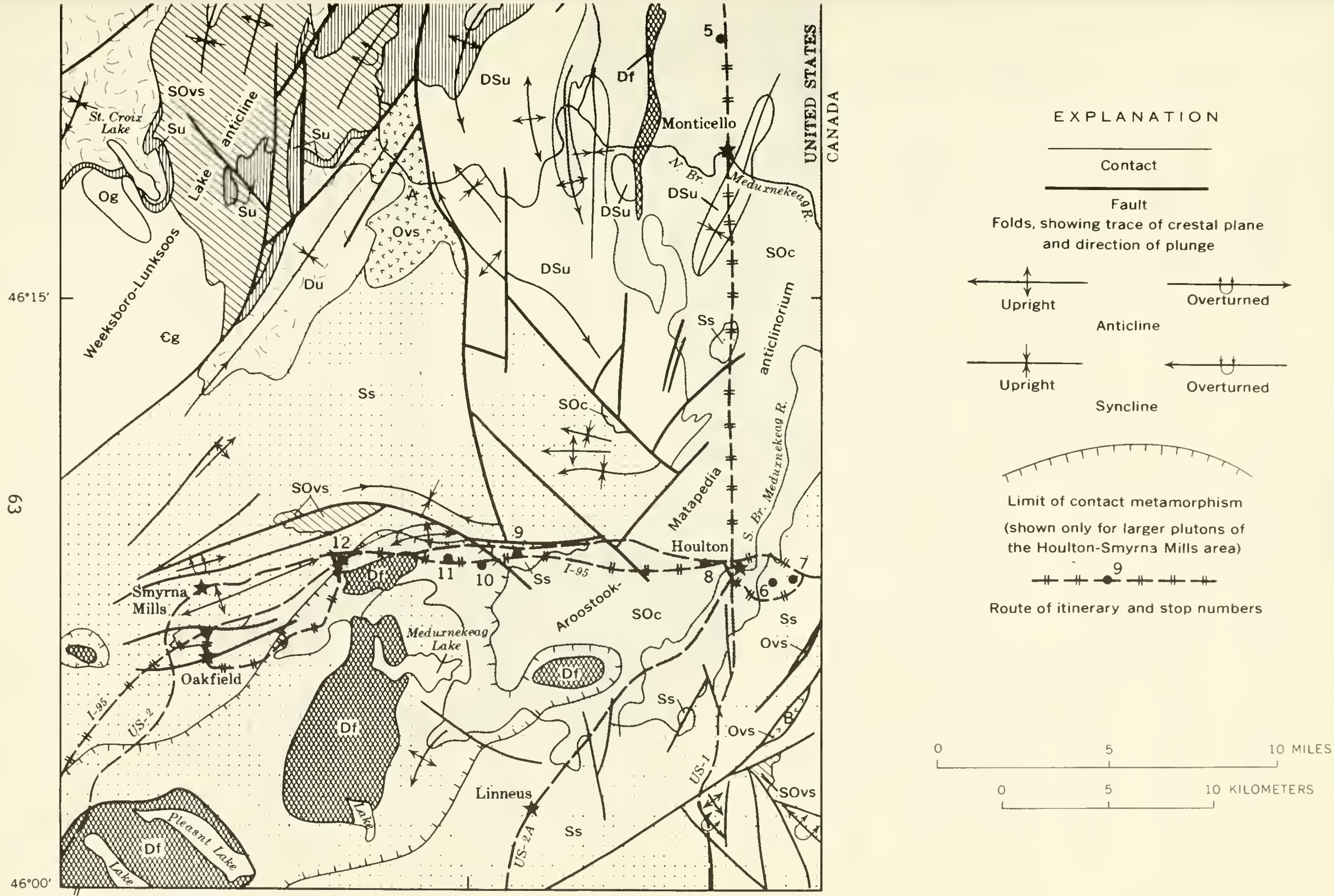
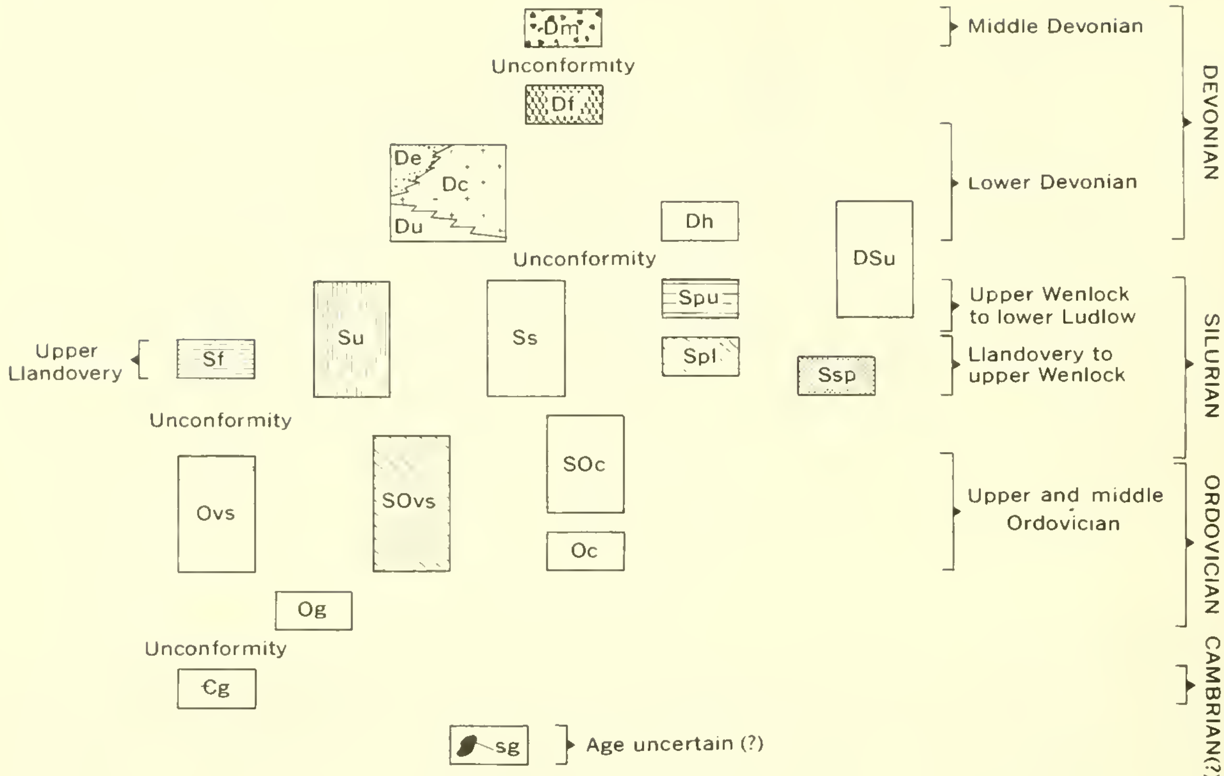


Figure 1. Generalized geologic map of northeastern Maine. Modified from Boucot and others (1964) and Pavlides (1965, 1968, 1971, 1972, and 1973). Points A,B,C, and D are referenced to figures 2 and 3.

EXPLANATION

Correlation of map units



Description of map units

Dm, MAPLETON SANDSTONE (MIDDLE DEVONIAN) -- Red and gray sandstone, conglomerate, and minor siltstone, ±2,200 ft (±670 m)

De, EDMUNDS HILL ANDESITE (LOWER DEVONIAN) -- ±4,000 ft (±1220 m)

Dc, CHAPMAN SANDSTONE (LOWER DEVONIAN) -- Greenish-brown graywacke and minor mudstone, ±8,000 ft (±2440 m)

Dh, HEDGEHOG FORMATION (LOWER DEVONIAN) -- Felsite, andesite, tuff, and interlayered sedimentary rocks, ±6,000 ft (±1830 m)

Du, UNDIVIDED ROCKS (DEVONIAN) -- Including:

(a) Swanback Formation (Lower Devonian) of the Presque Isle and Howe Brook quadrangles: greenish-brown sandstone, siltstone, and slate, ±10,000 ft (±3050 m)

(b) Seboomook Formation (Lower Devonian) of the Howe Brook quadrangle: cyclically layered, graded sets of gray-green sandstone, siltstone, and slate, +2,000 ft (+610 m)

(c) Unnamed gray slate and local conglomerate layers (Devonian) of the Howe Brook quadrangle, +2,000 ft (+610 m)

DSu, UNDIVIDED ROCKS (DEVONIAN AND SILURIAN) -- Including:

(a) Bell Brook Formation (Silurian? and Devonian) of Bridgewater and Houlton quadrangles: consists of an upper slate member, ±6,000 ft (±1830 m) and a lower conglomerate member, ±5,000 ft (±1525 m)

(b) Unnamed rocks (Silurian and Devonian) of the Houlton quadrangle: gray and green slate, siltstone, and conglomerate lenses, +2,000 ft (+610 m)

Spu, UPPER MEMBER OF PERHAM FORMATION (SILURIAN) -- Orange-weathered calcareous siltstone, shale, and slate with local limestone breccia lenses, ±2,000 ft (±610 m)

Spl, LOWER MEMBER OF PERHAM FORMATION (SILURIAN) -- Gray, green, and red shale and slate with local lenses of manganiferous and ferruginous shales and ironstones, +4,000 ft (+1220 m). Thickness from Roy (1970)

Ssp, SPRAGUEVILLE FORMATION (SILURIAN) -- Greenish calcareous siltstone and limestone, generally with bioturbation features, 0 to ±3,000 ft (±915 m)

Sf, FRENCHVILLE FORMATION (SILURIAN) -- Greenish graywacke and conglomerate, 0 to 4,000 ft (0 to 1220 m). Thickness from Roy (1970)

Ss, SMYRNA MILLS FORMATION (SILURIAN) -- Gray and green siltstone, quartzite, quartz graywacke, slate, phyllite, minor black slate and red and maroon siltstone, local manganiferous and ferruginous slate and ironstone layers, and local conglomerate and grit lenses, +6,000 ft (+1830 m)

Su, UNDIVIDED ROCKS (SILURIAN) -- Including:

(a) Maple Mountain Formation (Silurian) of the Howe Brook quadrangle: gray and green slate with sparse layers of quartzite and graywacke and local lenses of manganiferous slate and ironstone, ±6,000 ft (±1830 m)

(b) Burnt Brook Formation (Silurian?) of the Howe Brook and Bridgewater quadrangles: gray and green slate with sparse limestone layers, ±5,000 ft (±1525 m)

(c) Unnamed conglomerate and graywacke units (Silurian) of the Howe Brook quadrangle infolded and along the north and east margins of the Weeksboro-Lunksoos Lake anticline, 0 to 1,500 ft (0 to 460 m)

(d) Unnamed slate, phyllite, limestone, tuff, aphanite graywacke, and conglomerate (Silurian) of the Howe Brook quadrangle along west flank of Weeksboro-Lunksoos Lake anticline, ±1,600 ft (±490 m)

(e) Unnamed green and gray quartzite and slate (Silurian) of the Howe Brook and Mars Hill quadrangles, +4,000 ft (+1220 m)

SOc, CARYS MILLS FORMATION (MIDDLE ORDOVICIAN TO LOWER SILURIAN) -- Grayish-blue impure limestone with slate interbeds and local lenses of slate and graywacke, ±1,500 to +12,000 ft (±460 m to +3660 m)

SOvs, UNDIVIDED ROCKS (SILURIAN AND ORDOVICIAN) -- Including:

(a) Dunn Brook Formation (Ordovician or Silurian) of the Howe Brook and Bridgewater quadrangles: felsic and mafic metavolcanic rocks that include keratophyres, rhyolites, breccias, and basaltic and andesitic rocks, with local interbeds of slate and graywacke; for purposes of this report the unit also includes Spruce Top Greenstone, ±8,000 ft (±2440 m)

(b) Nine Lake Formation (Ordovician or Silurian) of the Howe Brook quadrangle: gray-green slate and minor purple slate, local lenses of volcanic rocks and conglomerate, ±4,000 ft (±1220 m)

(c) Unnamed green phyllite and slate (Silurian and Ordovician) of the Smyrna Mills and Houlton quadrangles, ±2,000 ft (±610 m)

(d) Conglomerate and siltstone (Silurian and Ordovician) at Mars Hill and nearby areas of uncertain thickness

Oc, CHANDLER RIDGE FORMATION (ORDOVICIAN) -- Green slate and graywacke with lesser amounts of siltstone and quartzite, ±5,000 ft (±1525 m)

Ovs, UNDIVIDED ROCKS (ORDOVICIAN) -- Including:

(a) Pyle Mountain Argillite of Ashgillian age (Upper Ordovician) of the Castle Hill anticline, ±600 ft (±185 m)

(b) Unnamed unit of Caradocian age (Ordovician) in the Castle Hill anticline: consists of felsic and intermediate volcanic rocks, chert, argillite, and black fissile shale, ±4,000 ft (±1220 m)

(c) Unnamed unit (Ordovician) in the Howe Brook quadrangle: consists of graywacke and black slate with local lenses of volcanic rocks and carbonaceous chert and argillite, +2,000 ft (+610 m)

(d) Unnamed unit (Ordovician) in the Houlton quadrangle: consists of black and gray slate of uncertain thickness

Єg, GRAND PITCH FORMATION (CAMBRIAN?) -- Slate, phyllite, quartz graywacke, quartzite, siltstone, and local lenses of metavolcanic rock, +2,000 ft (+610 m)

Df, INTRUSIVE ROCKS (DEVONIAN) -- Undivided felsic intrusive rocks (Devonian) that include:

(a) Granitic plutons of the Houlton and Smyrna Mills quadrangles

(b) Rhyodacitic dike and related felsite of Sugar Loaf Mountain in the Bridgewater quadrangle

(c) Garnetiferous granitic plutons of the Bridgewater quadrangle

(d) Granitic plutons of the Mars Hill and Presque Isle quadrangles

Og, Granitic pluton (Ordovician) of the Howe Brook quadrangle

sg, Serpentinite and metagabbro (age uncertain)

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with the Aroostook-Matapedia anticlinorium. The Carys Mills Formation and its equivalents were deposited on earlier Ordovician sediments, probably in a relatively deep-water environment where, in part, euxinic deposits were accumulating. This is suggested by the sulfidic black slates that contain radiolarian chert horizons with argillite partings that in turn contain British Zone 12 graptolites (Climacograptus bicornis zone), on the northwest side of the Aroostook-Matapedia anticlinorium (Pavlides, 1968, p. 24-25; 1971) and by the black and gray graptolitic slates on the southeast side of the anticlinorium (Pavlides. 1968, p. 25-26), shown respectively as points A and B on figure 1 (see also fig. 2). The Castle Hill anticline (fig. 1) contains similar rocks of this age. Zone 12 graptolites in black slates with and without associated volcanic rocks are widespread in the northern Appalachians (Berry, 1968, p. 28-30; Neuman, 1968, p. 41). These rocks probably were deposited in an archipelago environment characterized by volcanic island chains separated by deep-water troughs. The ancestral Carys Mills basin probably originated as an elongate trough in such an archipelago. The Chandler Ridge Formation (fig. 1), which consists of slate and graywacke without volcanic beds and which locally directly underlies the Carys Mills (Pavlides, 1968, pl. 1), is probably in part coeval with some of the Ordovician part of the Carys Mills and may represent a lenticular deposit shed laterally into the elongate Carys Mills basin from nearby land areas and islands.

By Llandovery time, in the Early Silurian, the Carys Mills basin was flanked by shallower shelf deposits and land masses (Ayrton and others, 1969, fig. 1; McKerrow and Ziegler, 1971, fig. 3). The faunal assemblages of the upper Llandovery rocks of this basin also support the interpretation of a basin with shallowing facies on its margins. This configuration of the Carys Mills basin probably became established as areas peripheral to the basin were uplifted during the Taconic orogeny (Pavlides and others, 1968); such areas shed clastic material into the basin to change the regimen of sedimentation. This event is recorded in the transition from the Carys Mills limy rocks to the less limy to nonlimy rocks of the Smyrna Mills Formation (Stop 12) and indicates uninterrupted sedimentation in the Carys Mills basin during Taconian time (Pavlides and Berry, 1966, p. B60). The more widespread areal distribution of the Smyrna Mills Formation and its temporal equivalents also suggests that the basin of deposition was probably wider at this time than during the time at which the Carys Mills Formation was deposited. A turbidite environment of sedimentation persisted in the Houlton-Smyrna Mills area of the basin, as indicated by such features as flute markings and convolute bedding present in the clastic rocks of the Smyrna Mills Formation (Stops 7 and 9).

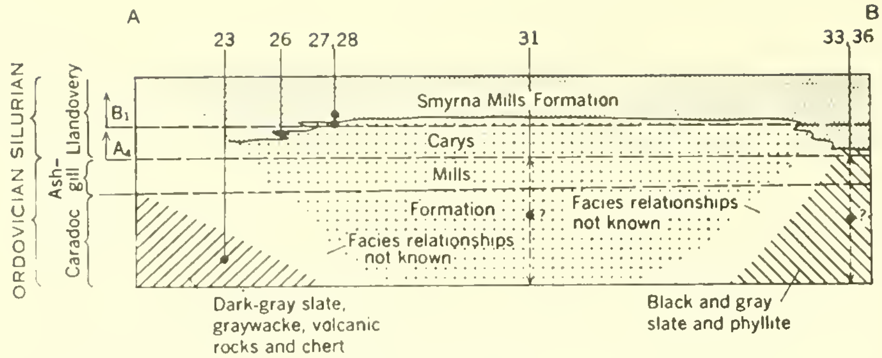

Figure 2. General facies relationships between Ordovician and Silurian rocks in the Carys Mills basin of the Howe Brook and Houlton quadrangles (see fig. 1 for location of points A and B). Numbers and leaders indicate fossil locality numbers that are keyed to figure 2 and plate 1 of Pavlides (1968), from which this figure has been modified.

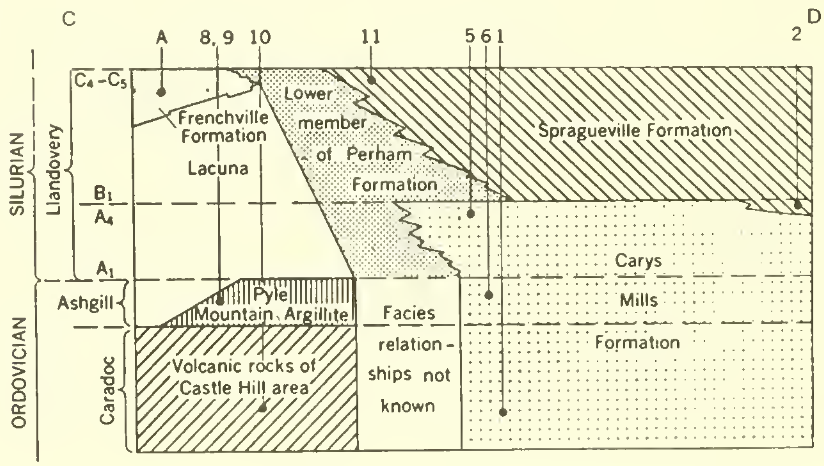

Figure 3. General facies relationships between Ordovician and Silurian rocks in the Carys Mills basin between the Castle Hill and Ft. Fairfield areas (see fig. 1 for location of points C and D). Numbers and leaders indicate fossil locality numbers that are keyed to figure 3 and plate 1 of Pavlides (1968) from which this figure is taken.

The lower member of the Perham Formation conformably overlies the Carys Mills Formation northwest of Presque Isle (see fig. 3 and Pavlides, 1968, pl. 1). Elsewhere in the Presque Isle area and near Ft. Fairfield, the Carys Mills is overlain gradationally by the Spragueville Formation (Stop 3), which consists chiefly of calcareous siltstone with abundant bioturbation features (Roy, 1970, p. 231-233), suggesting that probably a more quiescent and shallower but nonetheless offshore environment (see fig. 3) characterized the Carys Mills basin here in Early Silurian time. In the central part of the Presque Isle quadrangle, the Spragueville is conformably overlain by the shales and slate of the lower member of the Perham Formation, which in turn is conformably overlain by the calcareous siltstones and slates of the upper member of the Perham that contains shelly faunas and graptolites of late Wenlock and early Ludlow age (Pavlides, 1968, p. 21-22). These Silurian rocks are overlain by the Lower Devonian (upper Gedinnian) rocks of the Chapman syncline (fig. 1). The hiatus that separates these Silurian and Devonian rocks spans approximately the latest Silurian (Pridoli) and the earliest Early Devonian (early Gedinnian) time and represents the local Salinic disturbance of northern Maine. The Salinic is mostly recognized in northern Maine as a nonsequence. However, near the base of the Lower Devonian Seboomook Formation at St. Croix Lake in the southwest part of the Howe Brook quadrangle, conglomeratic lentils contain fossiliferous upper Llandovery (Silurian) cobbles, which indicate that local uplift also may have accompanied the Salinic disturbance (Pavlides, 1973).

Early Devonian volcanism, as represented by the Hedgehog Formation (fig. 1), took place after the Salinic disturbance in the Presque Isle area along the west margin of the Carys Mills basin, which, as a result of infilling, probably ceased to be a site of major sedimentation. Indeed, in Maine, Lower Devonian marine deposits occur only on the west side of the Aroostook-Matapedia anticlinorium (Pavlides, 1971, 1972, 1973, also fig. 1). After the volcanism that formed the felsic rocks of the Hedgehog Formation, a period of andesitic volcanism resulted in the deposition of the Edmunds Hill Andesite. This andesite, in part, supplied detritus to some of the marine sediments of the penecontemporaneous Chapman Sandstone (Stop 2). The nearby land areas that supplied debris into the Chapman basin also had an established flora by this time, as indicated by the <u>Psilophyton</u> debris locally found in the Chapman (Gregory <u>in</u> Williams and Gregoary, 1900, p. 133; Williams and Breger, 1916) and in the partly coeval Swanback Formation (Pavlides, 1973).

The Acadian orogeny, which probably occurred as a mid-Middle Devonian event in the northern Appalachians (Boucot, 1968, p. 92), deformed all the rocks of Early Devonian and older age in the region and destroyed the trough that the seaways occupied in this part of the Appalachians. Also, after the Acadian folding, some shallow-seated felsic plutons were emplaced, particularly in the Houlton-Smyrna Mills area (fig. 1). After Acadian deformation, the region emerged as a folded landmass on which local deposits of terrestrial origin, such as the Mapleton Sandstone (Stop 1), were laid down.

<u>Tectonics</u>.--Northern Maine (fig. 1) has undergone at least five tectonic events of varying intensities which have affected separate belts of rocks in different ways. The earliest deformation was that of the Penobscot orogeny (Neuman, 1967, p. 132) of Late(?) Cambrian to Early(?) Ordovician age (Hall, 1969), which folded and cleaved rocks of the Grand Pitch Formation of Cambrian(?)

age. These rocks are now exposed in the core of the Weeksboro-Lunksoos Lake anticline (fig. 1). Such rocks, in part, formed the terrane upon which were also deposited the volcanic rocks of the Dunn Brook Formation and the partly coeval slate, graywacke, and volcanic rocks of Graptolite Zone 12 time of the Middle Ordovician (which are presumed to be subjacent to the Carys Mills Formation (fig. 2)).

As the Carys Mills basin formed through subsidence, the terrane of the Weeksboro-Lunksoos Lake anticline became part of the western bounding shelf area that, in part, underwent Taconic deformation in Late Ordovician and Early Silurian time, as did some of the rocks within the Castle Hill anticline. The Carys Mills basin continued as a trough of uninterrupted sedimentation in Taconian time.

During and immediately after the Taconic event, the Weeksboro-Lunksoos Lake and Castle Hill terranes were the site of sandy deposition whose sparse shelly fossils, in places (Frenchville Formation), suggest nearshore brachiopod faunas that were redeposited with the offshore faunas at the base of relatively steep slopes (Berry and Boucot, 1970, p. 80, 102; Roy, 1973, p. 214-215). These sandy deposits may have formed at the base of a slope along islands or a paleoshoreline.

The Salinic event of approximately Pridoli to early Gedinnian time involved only local uplift and nondeposition. The Acadian orogeny, however, involved the most widespread deformation the area has undergone. All rocks up through the Lower Devonian were folded by this event, and a regional slaty cleavage formed. The earliest cleavage known, that of the Penobscot event, may have been transposed at this time into a shear cleavage that produced the passive-slip and passive-flow folds that characterize the Grand Pitch Formation in many places (Pavlides, 1973). This style of folding contrasts with the dominantly close folds of the flexure-slip and flexure-flow type found in younger rocks. Not all the terrane affected by Acadian deformation, however, was closely folded; the relatively broad and open folds of the Chapman syncline (fig. 1) reflect the more competent nature of the Chapman Sandstone and the underlying felsite of the Hedgehog Formation.

Elsewhere, where the rocks are argillaceous or limy, as in the Carys Mills Formation, or are thin-bedded sandstones containing argillaceous beds or manganiferous ironstones, the folding is more intense and complicated, doubly plunging folds being common. Inversion of plunge directions (Stop 5; Pavlides, 1962, pl. 5) also occurs locally. Disharmonic folding is present at places, such as that between the less competent rocks of the Silurian Maple Mountain Formation and the more competent beds of the underlying Silurian and Ordovician rocks in the Howe Brook quadrangle (Pavlides, 1973). The Silurian rocks beneath and close to the Chapman syncline may also be disharmonically folded with respect to the competent younger rocks of this fold. The competent rocks of the Chapman syncline, such as the Chapman Sandstone, are uncleaved (Stop 2), but, in part, this may reflect the weaker development of slaty cleavage in the northern part of the region, in contrast to the well-developed cleavage farther south. The absence of cleavage in the Spragueville Formation at Stop 3 probably reflects its tectonically sheltered position immediately beneath the competent rocks of the Chapman syncline.

The Houlton oroflex (Pavlides in U.S. Geol. Survey, 1971, p. A16-A17) is a sigmoidal flexure in the regional trend of folded rocks that occurs at about the latitude of Houlton, Maine (fig. 1). It probably formed during the late stages of the Acadian orogeny in this region but before the emplacement of the post-kinematic (uncleaved) felsic plutons that intrude the folded and cleaved rocks that the oroflex has sigmoidally warped. Some northward flexing of axial surfaces of folds along the Aroostook-Matapedia anticlinorium occurred locally, however, during the emplacement of some of these plutons (Pavlides, 1972, p. 3).

The Acadian regional slaty cleavage that locally grades into an incipient slip cleavage, for the most part, is parallel or nearly parallel to the axial surfaces of folds (Stops 7 and 8). Some of the northwest-trending folds northwest of Houlton, which formed at the same time as the Houlton oroflex, are cut at a large angle by this regional cleavage (Pavlides, 1965, pl. 1; 1971). The regional slaty cleavage along the Aroostook-Matapedia anticlinorium, however, is warped from a north- and northeast-trending direction (Stops 4, 5, 7, and 8) into a west trend (Stop 9) by the Houlton oroflex. Pelite has been injected into cleavage planes at several places in the Carys Mills terrane (e.g. Stop 8) and locally produces a false bedding effect (Pavlides, 1965, p. 52; 1971). A late west- to northwest-trending fracture cleavage, which in places cuts and crinkles the regional slaty cleavage (Pavlides, 1971, 1972, 1973), probably formed during the final stages of the formation of the Houlton oroflex. The Silurian-Devonian terrane northwest of Houlton also was probably faulted in large part as a result of the warping induced by movements related to the formation of the Houlton oroflex. The large generally north-trending fault in this general area may locally be a thrust (Pavlides, 1973).

The last deformational episode recorded in northeastern Maine involved the broad folding that affected the Mapleton Sandstone during the Maritime disturbance (Poole and others, 1970, p. 283, 295-296). The fault that bounds the Mapleton on its east side (fig. 1) may be a late event of this disturbance or even younger.

The mechanics by which the Houlton oroflex formed are uncertain. The oroflex could have formed as a result of a local shear couple operating on opposite sides of the Aroostook-Matapedia anticlinorium, the southeast block moving southwestward relative to the northwest block. Alternatively, northwest-southeast compression could have produced a similar warping, a mechanism suggested by Osberg (1968, p. 5) to explain the youngest northwest-trending folds of the Waterville-Vassalboro area of Maine. If the Acadian orogeny represents stresses generated by collision of continental margins resulting from the closing of the Proto-Atlantic, as suggested by Bird and Dewey (1970), either a shear couple or a northwest-southeast compressive stress could have been generated in central Maine, depending on the impingement angle of the European plate to the North American plate. An alternative to the plate-tectonics model was suggested by Chidester and Cady (1972) and can also be adapted to explain structures such as the Houlton oroflex and the Carys Mills basin. According to their suggestion, an oceanic ridge-rift system underlay the eastern North American continent. Beginning in Precambrian time, lateral spreading on either side of the rift system resulted in thinning of the sialic crust and concomitant downwarping of its surface to form the Appalachian eugeosyncline. Southeast of Newfoundland, along the Appalachian eugeosyncline, ultramafic rocks derived by upwelling from the mantle along the ridge-rift system were tectonically emplaced through the eugeosynclinal rocks that lay on an ensialic basement.

If this model is applicable, the Carys Mills basin may have formed as a result of a local and distinctly more active rift-spreading ridge subsystem within the Appalachian eugeosyncline. This subsystem should have been activated in about Caradocian time (Graptolite 12 and 13 intervals) to produce subsidence that formed the Carys Mills basin. Such a subsystem would have continued into the Acadian epoch, basin subsidence probably diminishing by Late Silurian time. The small masses of serpentinite and related rocks about 7 miles west-northwest of Bridgewater (fig. 1; and Pavlides, 1962, p. 28-30; 1965, pl. 1) may have been derived from such a subjacent rift-ridge system. In this model, the Houlton oroflex and related features elsewhere in Maine may have formed in response to transform fault displacements in the subjacent rift-ridge system or subsystems in late Acadian time.

None of the above hypotheses satisfies all the field relationships related to the Houlton oroflex as well as the general tectonics of the area. Perhaps additional ideas will be supplied by participants of this field trip. In any event, such speculations may serve to stimulate conversations that help pass the time during the long drive back to Orono from the field area.

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Itinerary

Assemble at Stop 1. To reach Stop 1 proceed to the center of Presque Isle and, at the traffic light at the intersection of U.S. Route 1 and Route 227 (immediately north of the Northeastland Hotel), turn west onto Route 227 and proceed 0.2 mile to junction with Route 163. Take Route 163 westward toward Mapleton for an additional distance of 2.2 miles to the intersection with Griffin Ridge Road (Route 7364). Turn north (right) onto Griffin Ridge Road and park along the east shoulder of the road, making sure you do not block the entrance to the town incinerator. Stop 1 is along Route 163 immediately west of the intersection with Griffin Ridge Road.

Mileage

0.00 Stop 1. Outcrops of the Mapleton Sandstone on the north side of the road. These outcrops are near the base of the formation at the southeast margin of the Mapleton syncline, a dish-shaped fold bounded by a fault along its east flank. The outcrops here are composed of sandstone and polymitic conglomerate which has yielded at least two fossiliferous cobbles containing brachiopods characteristic of the stratigraphically older Chapman Sandstone. The upper part of the Mapleton has more green-colored beds and, at one place, has yielded fossil plants of late Middle Devonian age.

The contact between the Mapleton Sandstone and underlying volcanic and marine sedimentary formations is not exposed. However, the mapped relationships indicate that it rests with angular unconformity on rocks of Early Devonian and older age. The unconformity here is ascribed to the Acadian orogeny.

View to the south is approximately along the axis of the Chapman syncline; a broad fold that is floored and rimmed by Lower Devonian sedimentary and volcanic rocks which probably overlie Silurian rocks unconformably; this unconformity represents the Salinic disturbance, which in northern Maine spans latest Silurian and earliest Devonian time.

Ridge in the immediate foreground is Hobart Hill and is held up by the Edmunds Hill Andesite which is coeval with the Chapman Sandstone, both of which overlie the volcanic rocks of the Hedgehog Formation. Hedgehog Formation rocks form the north-trending ridges that rim the Chapman syncline. Squa Pan Mountain (formerly Hedgehog Mountain) is the rimming mountain range on the west, and Quaggy Joe with Green Mountain on its south side are the ridges of Hedgehog Formation along the east flank of the Chapman syncline.

Return to cars and proceed north along Griffin Ridge Road.

1.45 Road junction. Turn south (left) and return to Route 163. Skyline ridge to south is Saddleback Mountain in the Howe Brook quadrangle; it is held up by volcanic rocks of the Dunn Brook Formation of probable Ordovician age. Lowland in the immediate foreground is underlain by the Chapman and Swanback Formations of Early Devonian age, within the Chapman syncline.

2.20 Turn east (left) and return to Presque Isle.

5.35 Junction of Routes 227 and 163. Turn right and continue along Route 163.

5.55 Junction of Routes 163 and U.S. Route 1 at traffic light. Turn right and continue through town to Chapman Street.

5.70 Junction U.S. Route 1 and Chapman Street. Turn right onto Chapman Street and follow road southwestward through outskirts of town.

6.05 Railroad crossing; Presque Isle Stream to west.

7.35 Quaggy Joe Mountain on skyline to east is held up by volcanic rocks of the Lower Devonian Hedgehog Formation along the east flank of the Chapman syncline.

8.00 Road junction; continue southward on Chapman Road.

9.00 Squa Pan Mountain on skyline to the southwest is along the west flank of the Chapman syncline and is also held up by volcanic rocks and minor amounts of sedimentary rocks of the Hedgehog Formation.

10.80 Turn cars around on farm road and park along east side of Chapman Road facing to the north.

Stop 2. Outcrops of the Chapman Sandstone on the east side of the road. The Chapman grades southward into the stratigraphically lower Swanback Formation, and to the north it interdigitates with volcanic and volcaniclastic rocks of the Edmunds Hill Andesite. Brachiopod faunas from the Chapman and from an ash bed in the Hedgehog are of Early Devonian (early Helderbergian) age. The Chapman, Swanback, and Hedgehog Formations are pre-Acadian and unconformably underlie the Mapleton Sandstone. The broad and open nature of the Acadian folding that characterizes the Chapman syncline is believed to be related to the stiffening effect of the competent sandstone and volcanic rocks that make it up. This style of folding contrasts with the closer folding of the less competent beds in the subjacent Silurian rocks which are characterized by steep to vertical dips.

The Chapman Sandstone commonly consists of fine- to medium-grained massive sandstone beds as much as 4 feet thick containing minor beds of shale or mudstone. Quartz and chloritic volcanic detritus are the chief constituents of the sandstones, and plagioclase and muscovite are present in minor amounts; chloritic volcanic detritus is most abundant in the Chapman Sandstone near Edmunds Hill, where the Chapman appears to interdigitate with the Edmunds Hill Andesite.

Return to cars and proceed along Chapman Road.

12.10 Crossing Hughes Brook.

13.65 Junction with road leading to Aroostook State Park. Turn southeast (right) and proceed toward Spragueville.

14.80 Mars Hill monadnock on skyline.

15.60 Possible Stop 2A. Graptolite fossil locality in upper member of the Perham Formation approximately 500 feet below the base of the Hedgehog Formation. Monograptid collections from this locality and elsewhere from this formation are of the zone of *Monograptus nilsonni* of early Ludlow (Late Silurian) age. These fossils and stratigraphic relationship elsewhere in the

Presque Isle area indicate that the age span of the upper member of the Perham is late Wenlock to early Ludlow. The hiatus between the Lower Devonian Hedgehog Formation and the upper member of the Perham Formation of late Wenlock to early Ludlow age represents the Salinic disturbance of northern Maine.

16.10 Community of Spragueville; continue east to the top of the hill.

16.35 Stop 3. Outcrops of Spragueville Formation along the south side of the road.

The Spragueville is generally a gray calcareous siltstone that weathers to a pale green. The bedding surfaces are generally rough and may be planar or gently warped. Fine-scale internal laminations are characteristic but commonly destroyed or difficult to see because of the rather extensive bioturbation present in most of these rocks.

On the basis of the age of graptolites, ostracodes, and brachiopods found in the Presque Isle and Ft. Fairfield areas, the Spragueville spans about A_1-B_4 to C_3-C_5 Llandovery time. Because at Spragueville the lower member of the Perham is absent and the Spragueville directly underlies the upper member of the Perham (late Wenlock to early Ludlow in age), the Spragueville may extend upward into middle Wenlock age, or the upper member of the Perham here may extend downward into the late Llandovery. Alternatively, the Spragueville and the upper member of the Perham may be separated by a disconformity in this area.

The abundant bioturbation that characterizes the Spragueville is absent in the other Paleozoic rocks of northern Maine, including formations coeval with the Spragueville, such as parts of the lower member of the Perham Formation and the Frenchville Formation (see fig. 3). The Spragueville appears to be a more offshore deposit than these latter formations. It directly overlies the Carys Mills Formation which is, for the most part, a calcareous flysch which in places contains abundant turbidite features. Apparently the Spragueville was deposited within the Carys Mills basin in a local, relatively quiescent environment having a low rate of deposition and abundant bottom-feeding and burrowing animals. Internal grazing tracks of deposit feeders are common and are primarily responsible for the bioturbation.

17.10 Junction with U.S. Route 1. Turn south (right) onto U.S. Route 1 and proceed south toward Mars Hill.

19.25 Crossing Clark Brook

20.85 Mars Hill on skyline to the southeast

24.15 Crossing Young Brook

24.45 Crossing town line of Mars Hill

25.15 Turn right onto dirt road.

25.30 Stop 4. Road-metal quarry in Carys Mills Formation which here consists of thin- to medium-bedded ankeritic and calcic impure limestone interlayered with slate and less commonly with quartzite layers. Limestone layers locally show grading and some crossbedding and in a few places convolute bedding.

Steep to vertical bedding and local dextral (dominant) and sinistral folds with steep to vertical plunges are visible in various parts of the quarry. Note the absence of "drag folds" in the vertical plane. Based on sedimentologic features, tops generally face southwest. Slaty cleavage is well developed in pelitic layers, but cleavage is more widely spaced in limestone beds.

Return to cars and proceed southeastward along dirt road.

26.90 At community of Blaine and intersection of U.S. Route 1. Turn right onto U.S. Route 1 and proceed south toward Bridgewater. Route from Blaine southward to Stop 5 will be along folded rocks of the Carys Mills Formation. Bedding and cleavage throughout this region are steep to vertical.

27.60 Skyline hills to southwest are held up by metavolcanic rocks of the Dunn Brook Formation. Ridge with fire lookout tower is No. Nine Mountain in the Howe Brook quadrangle.

28.20 Outcrops on east side of road are folded and cleaved calcareous siltstones and impure limestones interbedded with slate layers of the Carys Mills Formation. Convolute bedding is locally well developed; sweat veinlets of calcite cut the rocks.

28.90 Crossing Three Brooks

29.10 Outcrops on east side of road of vertically dipping limy beds of the Carys Mills Formation with subhorizontal kink bands.

29.90 Outcrops on east side of road of Carys Mills Formation with nearly vertical dips and an incipient slip cleavage that has produced a small fold. The top of the hill to the east is held up by a lens of slate within the Carys Mills Formation.

30.60 At Bridgewater Corners. Hill to the west, the north slope of which has been cleared for farming, is held up by a garnetiferous granitoid rock that has intruded and thermally metamorphosed the limy rocks of the Carys Mills Formation.

32.10 Crossing B & A Railroad tracks in Bridgewater

33.10 Outcrops in the field immediately west of U.S. Route 1 show wavy-bedded Carys Mills Formation with differential weathering between limy and less limy beds, which produces an etched, ribboned surface and which led to the usage of "ribbon rock" in some of the early reports of this area. Small conical hill on skyline to the southwest is Sugar Loaf. It is held up by felsic volcanic rocks that occur at the north end of a 5-mile-long rhyodacitic dike.

35.80 Subdued outcrops of Carys Mills Formation; note moderate (40°) dip of bedding to the northeast.

36.10 Stop 5. Road-metal quarry in calcareous flysch of the Carys Mills Formation. Bedding and cleavage dip steeply. Steep dips of bedding define an antiformal flexure that is broken locally by a fault. Fold has northeast-trending axial plane, and mullion structure is present near the hinge area of the fold. Note abundant calcite sweat veinlets that cut the rocks.

Sedimentary structures indicate, however, that the steeply dipping beds in the antiform are overturned so that in reality the fold is the inverted nose of a stratigraphic syncline. Sedimentary structures visible in the beds are convolute and graded bedding. Flame structure in convoluted layers is also visible. In general, convoluted beds consist of peaked anticlines and more rounded or open synclines; the peaked anticlines point in the top-facing direction of bedding.

36.80 Mullion structure in Carys Mills Formation on west side of roadcut.

37.60 Jewells Corner

39.40 Crossing North Branch of the Meduxnekeag River in Monticello

44.05 Outcrops on east side of road of Silurian Smyrna Mills Formation that occur in an infolded syncline within the Carys Mills terrane. Smyrna Mills Formation here consists of quartz graywackes, micaceous quartzites, slate, and interbeds of limestone like that of the Carys Mills Formation. Presence of such limestone near the base of the Smyrna Mills is common, as the contact between the Smyrna Mills and the underlying Carys Mills is gradational. Many of the sandstone beds of the Smyrna Mills here show graded bedding, crossbedding, climbing ripples, and convoluted layers. Good bedding-cleavage relationships are also present.

44.50 Community of Littleton

44.55 Outcrops on east side of road of Smyrna Mills Formation as at mileage 44.05. Here, however, thick (4 to 5 feet) beds of quartz graywacke are present. Some beds display well-developed flute casts and channel scours.

44.80 Littleton Ridge is on skyline to the west. It is underlain by slate, graywacke, and grit of Silurian and (or) Devonian age. Some of these rocks are probably a facies of the Smyrna Mills Formation.

45.00 Highlands to the southeast are held up by clastic rocks of the Smyrna Mills Formation. Such rocks generally form uplands on either flank of the Aroostook-Matapedia anticlinorium which is mostly cored by the less resistant limy rocks of the Carys Mills Formation.

45.90 Crossing Big Brook

48.80 Maine State Police barracks on west side of road

50.20 I-95 interchange

50.65 Turn left onto Washburn Street - Texaco gas station on northeast corner. Go one block.

50.70 Turn right from Washburn onto Highland Ave.

50.95 Bridge over South Branch Meduxnekeag River

51.00 Turn left onto Pleasant Street

51.80 T intersection. Turn right and go one block to T intersection with Route 2. Turn left onto Route 2.

52.20 Turn left at Hancock Barracks sign and proceed to Fort.

52.50 Stop 6. Lunch at Hancock Barracks

Hills to west in foreground are the Oakfield Hills which are mostly hornfels of the Smyrna Mills Formation that rim the granitic Hunt Ridge and Pleasant Lake plutons. On a clear day it is possible to see Mt. Chase (Ordovician volcanic rocks) and Mt. Katahdin (Devonian quartz monzonite).

Ridge to the south is Westford Hill which consists of Smyrna Mills Formation with essentially northwest-trending folds that contrast with northeast-trending folds of the region.

Hancock Barracks were built and manned by the U.S. soldiers who were sent to Houlton during the border dispute between the U.S. and Great Britain in 1839 during the bloodless so-called Aroostook Wars.

After lunch return to Route 2 and turn northeast (left) and proceed towards Canadian Border.

52.80 Outcrop on right side of road is Smyrna Mills Formation. Skyline ridge is along International Boundary line with Canada. Distant hill to the north is south end of Mars Hill.

53.05 <u>Stop 7</u>. Park along shoulder of Route 2 and enter quarry on north side of road on foot.

Smyrna Mills Formation here consists of thinly bedded gray-green micaceous quartzite and quartz graywacke, thick-bedded quartzitic beds, and thin beds of slate.

The northwest end of the quarry exposes a tight asymmetric anticline plunging 45° NE. The core of the fold has thin-bedded layers that are overlain by thick-bedded quartzite on its flanks. Load casts and associated flame structures and other sole markings on the bottom of thick-bedded quartzites show normal top-facing directions and indicate that fold is both a structural and stratigraphic anticline. Slaty cleavage is present in thin-bedded part of rocks but is not as conspicuous in the heavy-bedded layers.

Southwest part of quarry also has northeast-plunging anticline, as outlined by thick-bedded quartzites. Well-developed flute casts on the southeast limb also indicate that structure is right-side-up and that the local current that formed these flute markings flowed approximately westward.

Ludlow-age graptolites (scarce) have been found in a few rock fragments within this quarry, and a little crinoidal debris is also present. Trace fossil tracks of spiraling meanders have also been noted on some slabs.

Return to cars and continue northeast toward Canadian Border.

53.25 Outcrops of thin-bedded Smyrna Mills Formation on southeast side of the road with a 1-inch-thick ash layer.

53.30 Crossing Cook Brook; stay in left lane for turn onto I-95.

53.85 Take U-turn ramp onto I-95 and stay on I-95 heading westward toward Houlton.

54.65 Crossing infolded Carys Mills Formation near the contact with the Smyrna Mills Formation. Lowland area in the immediate foreground is underlain by limy rocks of the Carys Mills Formation; upland area on the skyline is held up by Smyrna Mills Formation on the northwest flank of the Aroostook-Matapedia

anticlinorium, which to the north has been trending north and northeast, is here warped into a westward trend by the sigmoidal flexure of the Houlton oroflex. Both bedding and the regional slaty cleavage are rotated from north to west trends.

In the region about 8 to 15 miles northwest of Houlton, the rocks are folded along northwest-trending axes, and much of this and the nearby terrane are broken by inferred faults. These northwest-trending folds and the associated faults reflect the deformation induced at the time the Houlton oroflex began to form.

55.05 Crossing gradational contact between the Smyrna Mills and Carys Mills Formations. Hills on the skyline to the southwest are underlain by hornfels that rims the granitic plutons in the Oakfield Hills area.

55.45 Crossing Meduxnekeag River

56.55 Crossing B & A Railroad tracks

56.75 Crossing B Stream

56.95 Stop 8. Park on shoulder of road. Carys Mills Formation has open folds that are cut by cleavage and that plunge about 30° SSW. in general conformity with the plunge direction of the Aroostook-Matapedia anticlinorium here. Folds best seen on south side of roadcut. Outcrops in the island between the divided highway at the authorized U-turn road show pelite injection along cleavage that imparts a cleavage banding or "false bedding" to the rocks. Because of the gentle dip of the beds, some of the bluish calcic layers have been preserved between the cleavage slices and further accentuate the false bedding effect.

57.15 Crossing anastamosed esker which merges southward to become a single esker through the Houlton quadrangle. Carys Mills outcrop beneath eastern flank of esker has yielded a single graptolite of possible Ordovician age.

57.45 Carys Mills Formation with gently plunging folds.

58.05 Carys Mills Formation on north side of road. From here to Stop 9, I-95 is along west-trending limb of Houlton oroflex and mostly through rocks of the Carys Mills Formation.

58.75 Crossing Moose Brook

59.05 Crossing New Limerick town line

60.25 Crossing Ludlow town line

60.75 Hills on skyline to the south are composed of hornfels of the Smyrna Mills Formation that rim the Hunt Ridge pluton.

61.45 French Road overpass. Smyrna Mills Formation synclinal inlier. Smyrna Mills outcrops are exposed on the north side of the road immediately east of the French Road overpass.

62.35 Stop 9. Tight upright syncline within Smyrna Mills Formation that plunges about 20° E. I-95 is here located on synclinal axis. These outcrops are also along the west-trending part of the Houlton oroflex.

Beds on the south limb of the syncline (south side of road) have structural (small folds and bedding-cleavage) and sedimentary (ripple marks, graded bedding, convolute bedding) features that prove way-up directions. A conglomerate in a separate syncline immediately north of Stop 9 is interpreted as having formed in a high-energy environment and having been shed as a lenticular deposit into a more offshore environment (Smyrna Mills seas) from an area to the north or northwest which was undergoing uplift in the late phases of the Taconic orogeny.

Immediately west of Stop 9, along I-95, high-grade calc-silicate rocks are contact metamorphosed limestones of the Carys Mills Formation. The absence of any observable thermal metamorphism in the Smyrna Mills rocks at Stop 9 suggests that a fault of postcontact metamorphic age separates the rocks of Stop 9 from the calc-silicate terrane to the west.

62.65 Crossing Mill Brook

63.05 Stop 10. Contact metamorphosed Carys Mills Formation with well-developed compositional layering (bedding). Contortions and grooves (lineations) probably formed at time of contact metamorphism concomitant with the emplacement of subjacent and nearby granitic rocks. Light-colored layers in places contain wollastonite-diopside-calcite-grossulerite and idocrase-wollanstonite-diopside assemblages. Dark layers (purplish gray) are generally biotitic and may contain diopside, cordierite(?), and calcic plagioclase. Elsewhere in less metamorphosed rocks, tremolite is common, and wollastonite and idocrase are absent.

63.15 Crossing Lamb Brook (topographic sheet designation), which the Maine Highway Commission has labeled as Haggerty Brook.

63.75 Crossing stream designated as Lamb Brook by Maine Highway Commission but which Smyrna Mills topographic map shows as an unnamed brook that drains southward into Bradbury Lake.

64.00 Calc-silicate rock of Carys Mills Formation

64.15 Stop 11. Folded wollastonite-idocrase calc-silicate rocks of the Carys Mills Formation. Note large idocrase metacrysts here. Wollastonite-idocrase-diopside-garnet metalimestone assemblages are present here.

65.05 Entering Smyrna township. Low-lying hill immediately to the south is held up by the Cockrane Lake pluton (hornblende-biotite microcline and microcline-perthite granite). High hills to the southeast are hornfels of the Smyrna Mills Formation, a hornfels that is more resistant to weathering than the calc-silicate rocks (Carys Mills metalimestones) around other plutons.

65.55 Outcrops here are lower grade calc-silicate rocks nearer the margins of the thermal aureole than the wollastonite-idocrase-bearing calc-silicate rocks of Stops 10 and 11.

65.75 Smyrna Interchange. Leave I-95 here. Rocks in roadcut are tremolitic calc-silicates of the Carys Mills Formation.

66.00 Intersection of exit ramp from I-95 and U.S. Route 2. Turn west (right) onto Route 2.

66.30 Ascending hill underlain by Smyrna Mills Formation

66.60 Subdued outcrops in ditch and field on the north side of road contain graywacke and grit mapped as part of a lentil near the base of the Smyrna Mills Formation. These sandy to gritty and locally conglomeratic rocks were probably redeposited from the shore of land adjacent to the Carys Mills basin that was undergoing Taconian uplift. Grit and sandy layers are locally graded and also channel into older beds. These features indicate that tops face northwest, consistent with the local stratigraphy and structure.

66.80 Outcrops of calcareous siltstone and slate with *Platyceras* and some fragmental brachiopods. Sparse pebbles occur in these rocks.

Although the Smyrna Mills Formation is characterized by graptolitic faunas, these shelly-bearing rocks may be allochthonous, redeposited from accumulates elsewhere in a nearshore environment.

66.90 Turn south (left) onto Route 6098.

67.25 Outcrops of Smyrna Mills Formation on the east side of road.

67.30 Crossing Limestone Brook

67.50 Stop 12. At overpass of I-95. Gradational contact between Smyrna Mills Formation and Carys Mills Formation. Grit-graywacke lens described for mileage station 66.60 crops out as small grit exposure in northeast corner of front yard of house immediately south of Limestone Brook. Graptolite faunas of the same Early Silurian age (British Graptolite Zone 12) occur on either side of the contact, and the regional conformity of the two formations indicates that here and elsewhere the Carys Mills basin did not undergo Taconian deformation.

The results of the Taconic orogeny in areas peripheral to this trough are reflected, however, in the change in sedimentation from the essentially limy rocks of the Carys Mills Formation into the more clastic, less limy rocks of the Smyrna Mills Formation.

Limy beds of the Carys Mills Formation that occur in a small warp immediately south of the overpass of I-95 contain convolute layering which indicates that tops face north toward the younger Smyrna Mills Formation.

67.70 Timony Mountain is in immediate foreground to the south; ridge to the southwest is Sam Drew Mountain. These ridges are underlain by biotite-cordierite hornfels of the Smyrna Mills Formation that rim the Hunt Ridge pluton south of Timony Mountain.

68.30 Crossing B & A Railroad tracks. Railroad cut west of the road along the tracks consists of steeply dipping Carys Mills Formation with abundant boudinage and pull-apart features These structures may have formed at the time the nearby Hunt Ridge pluton was emplaced.

68.90 Mt. Katahdin is on the skyline to the west. Peaked mountain to the north is Mt. Chase.

72.00 Crossing B & A Railroad tracks in Oakfield

72.30 T-intersection at west end of Oakfield. Turn north (right) toward junction with I-95.

72.90 Junction with I-95. Take I-95 south to Orono. Highly contorted beds of Carys Mills Formation are exposed at this interchange along I-95. About 3/4 mile to the west, the Carys Mills Formation, which here forms the core of the Aroostook-Matapedia anticlinorium, plunges beneath the Smyrna Mills Formation.

End of field trip. Continue along I-95 to Orono.

PRECAMBRIAN ROCKS OF SEVEN HUNDRED ACRE ISLAND AND DEVELOPMENT OF CLEAVAGE IN THE ISLESBORO FORMATION

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Introduction

Islesboro is a low-lying island, 10 miles long and nowhere more than 2 1/2 miles wide, with a deeply indented shoreline. This island, together with structurally related islands (see Fig. 1) that extend southward another 4 miles, will be referred to as the Islesboro block.

The field trip will focus on three geological features of the Islesboro block. The first is the stratigraphy and lithology of the chlorite-grade metamorphic rocks that compose most of the block and constitute the Islesboro Formation. These rocks are different in several aspects from those immediately adjacent on the surrounding mainland, and compose a structural block which I interpret to be bounded by large north-trending faults along which miles of right-lateral strike-slip motion has occurred. Details of the structural blocks of northern Penobscot Bay are given in trip B-7, this volume.

The second feature is a horst of Precambrian medium-grade metamorphic rocks within the Islesboro block. These rocks were the first of Precambrian age to be recognized in Maine and have many implications that bear on the nature of basement rocks along the northeastern Maine coast and of the faulting in the region.

The third feature is the relation between large faults and cleavage and foliation in chlorite-grade rocks. Outcrops are spectacular and should provide valuable background for geologists who will later work with deformed higher grade metamorphic rocks with chlorite-bearing protoliths.

The rocks to be seen were most recently mapped mostly by Paul C. Bateman and Harvey E. Belkin of the U.S. Geological Survey in 1968 as part of my mapping project on the Castine and Blue Hill 15-minute quadrangles. This map is reproduced (Fig. 1), with few changes except for the omission of hundreds of structure symbols. D.G. Brookins, now of the University of New Mexico, and I established the age of the Precambrian rocks. I am responsible for postulating boundary faults of large displacement on either side of Islesboro. Another interpretation based on the same map was given by Osberg in Poitras and others (1972).

The field trip will necessarily be hurried to meet the ferry schedule. Additional problems can be caused by a trip by charter boat necessary to see important outcrops on Spruce and Seven Hundred Acre Islands. Bad sea or tide conditions may force the order of stops to be changed or may prevent access to some stops. Swimming in Penobscot Bay in October can be hazardous to your health and is at best marginal in August.

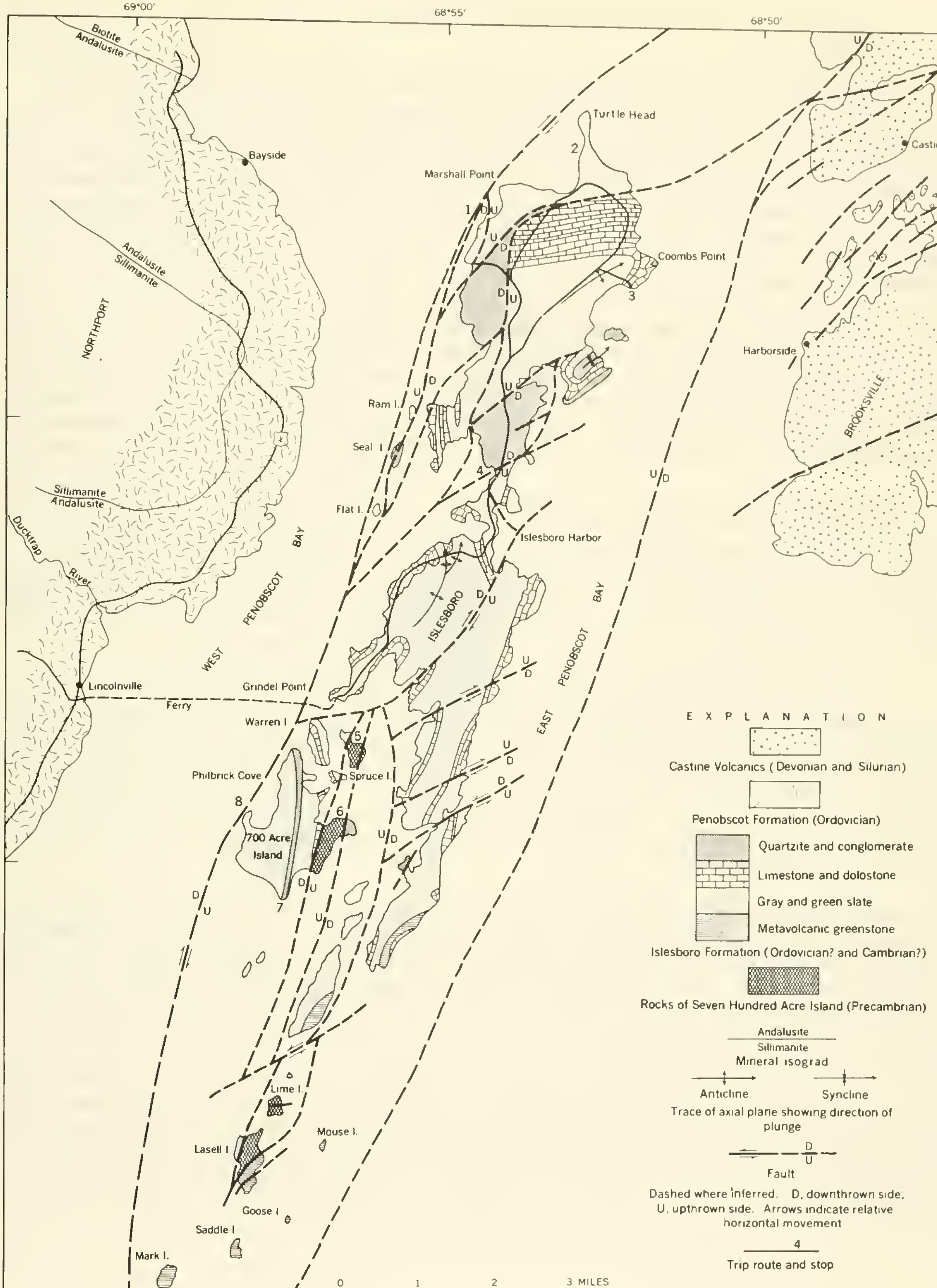

Figure 1. Geologic map of Islesboro and vicinity.

The Metasedimentary Rocks of the Islesboro Formation

A miogeosynclinal sequence of slate, carbonate rock, quartzite, and quartz conglomerate makes up all except the southernmost part of the Islesboro block. Sedimentary rocks seen on this trip on the island of Islesboro proper are conspicuously poor in feldspar, suggesting that deposition took place on a slowly subsiding platform that received only mature sediment. Thin and persistent bedding is still observable. Common graded and crossbeds indicate that most of the folded section is upright, and it would seem possible at first glance to reconstruct a sedimentary section and separately to name many members. This, unfortunately, is not possible because of many faults, lack of diagnostic marker beds, and an absence of fossils.

The southern point and most of the small islands south of it contain abundant feldspathic greenstones and greenschist, which become predominant, complete with pillowed lavas on Saddle, Mouse, and Goose Islands and Robinson Rock in the Vinalhaven quadrangle. The trip will not include a visit to the greenstone terrane, but its presence is noted here because of its relevance to attempts to correlate rocks of the Islesboro block with contiguous blocks.

My hypothesis is that the stratigraphic sections exposed on Islesboro have so few similarities with the stratigraphic columns in adjacent structural blocks (see trips A-4 and B-7) that very substantial fault motions must be postulated. Abundant evidence of faulting will be seen on the trip. My hypothesis differs from one postulated by Osberg and Guidotti, currently working on rocks with some similarities but mostly of very much higher metamorphic grade to the west and south in the Camden and Rockland areas (see trip A-4).

In 1956, I correlated the Ellsworth Schist with the Islesboro Formation, on the basis of similarities between a poorly exposed Ellsworth section at North Deer Isle, which contains a thick dolostone, and the greenstone-rich section of the southern part of the Islesboro block. However, the differences between the feldspar-rich Ellsworth sediments and most of the rock types on Islesboro discredit this correlation for most if not all of the two sections. Conceivably they could be contemporary sections deposited on the same platform at widely separated areas.

Earlier, Smith, Bastin, and Brown (1907) correlated the greenstones on Islesboro with the North Haven Greenstone, which is now thought to be equivalent to part of the Ellsworth Schist (Brookins, Berdan, and Stewart, 1973, p. 1621). Smith, Bastin, and Brown (1907, p. 3) also correlated a few hundred feet of sedimentary rocks on the east shore of Rockport Harbor to the south with some of the rocks of the Islesboro Formation exposed more than 10 miles northeast. The main criterion they used was clearly stated: "In view of the striking and unusual character of the quartzite conglomerate, it seems highly improbable that two occurrences so closely adjacent as those of Islesboro and the mainland can be of different age." All of the massive buff quartzite and clean quartzite conglomerate in the region was called Battie Quartzite after the Mt. Battie, Camden, locality on the mainland. The Penobscot Formation that makes up most

of the mainland west and north of Islesboro was thought to overlie the Battie Quartzite.

Smith, Bastin, and Brown (1907, p. 1-3) named the Islesboro Formation that part of the metasedimentary section on Islesboro which they thought lay beneath the quartzite. Their Coombs Limestone Member was almost immediately beneath the quartzite, and their slate member was beneath the Coombs Limestone Member. Stop 3 is the locality they cited to demonstrate these relationships. The slate member was recognized to be mostly slates, but also contained calcareous slates, impure quartzites, carbonate beds, and small amounts of pyroclastic rocks. The 1907 map and text discussion and the recent mapping have established several additional beds of dolostone and limestone that are not stratigraphically related to quartzite and several beds of quartz conglomerate lacking adjacent carbonates. These observations and the recognition of the many faults that break the Islesboro section assure that the correct stratigraphic section of the Islesboro Formation is complex and is not now known because of inability to correlate between fault blocks. Thus, the old correlation by Smith, Bastin, and Brown of the thin Rockport Harbor section with a small section of the Islesboro Formation must be questioned. Possibly the detailed work of Osberg in the Rockport area will clarify this issue.

The age of the Islesboro Formation is only known to be pre-Devonian and post-late Precambrian. On the basis of style of deformation and lithology it probably is Cambrian and (or) Ordovician. Rb/Sr whole-rock isochrons by Brookins (written communication, 1971) are compatible with an early Paleozoic age but have a large associated error band.

Representative sections of the Islesboro Formation were selected for stops on this trip. Most of them also show deformations related to faulting, but throughout it will be emphasized that the correlation of the Islesboro Formation is a problem with profound implications to the geologic interpretation of Penobscot Bay country.

Precambrian Rocks of Spruce and Seven Hundred Acre Islands

Outcrops of splendent coarse muscovite schist, garnet-bearing amphibolite, retrograded garnet-andalusite schist, quartzite, and banded gray marble containing muscovite pegmatite can be seen on Spruce, Seven Hundred Acre, Lime, and LaSell Islands in a horst; these medium-grade rocks contrast markedly with the chlorite-grade rocks of the rest of the Islesboro block and are nowhere in stratigraphic contact with the Islesboro Formation. Smith, Bastin, and Brown (1907, p. 9) speculated that the regional deformation might be younger than the pegmatite on Spruce Island because the muscovite it contains is highly deformed, and the pegmatite is remote from and different in composition from other granites in the region. They concluded, however, that deformation more probably resulted from differential flow during intrusion. K/Ar and Rb/Sr dating of the muscovite by Brookins and Stewart (written communication, 1972) proves that the pegmatite is, in fact, late Precambrian in age ($\sim 600 \pm 20$ m.y.) and older than the faulting that breaks the Paleozoic rocks of the Islesboro block. Rb/Sr whole-rock dating

with a large error band also indicates that the medium-grade metamorphic rocks are older than the pegmatite. The medium-grade rocks have been extensively retrograded and are more foliated than the pegmatite. No migmatite has been observed, and the deformation style is simple.

The closest known Precambrian rocks are the gneisses in the Passagassawakeag block 12 miles northwest of the Precambrian rocks of Islesboro. Mapping of these gneisses by Wones and Bickel (see trips A-1 and B-7) has shown three deformations older than Late Ordovician, multiple intrusion of granitic rocks, abundant migmatite, and higher metamorphic grade (to second sillimanite) throughout the Passagassawakeag block. The apparent contrast in Precambrian basement rocks in the Islesboro and Passagassawakeag blocks is one possible consequence of an early Paleozoic plate collision (Trip B-7).

The Precambrian rocks of Seven Hundred Acre Island have many similarities to the Green Head Formation Precambrian of southwestern New Brunswick. The finely laminated dolomitic limestone of Lime Island is virtually identical in appearance with dolomitic limestone at Green Head, St. John, New Brunswick. Thin beds of varied composition are common to both of these Precambrian terranes. The Precambrian rocks of Seven Hundred Acre and nearby islands are therefore tentatively correlated with the Green Head Formation. If the Precambrian rocks of the Islesboro block were originally contiguous with the Green Head Formation of New Brunswick, more than 100 miles of strike-slip motion would have been required to bring them to their present positions.

Development of Cleavage and Schistosity near Large Faults

The Islesboro block is bounded on both sides by major regional fault zones. Both fault zones are probably branches of a single fault zone, here called the Turtle Head fault zone; south of the Islesboro block this zone strikes nearly north; north of the block, it strikes N.35°E. Condidered in this way, the entire Islesboro block is a fault sliver caught in a bend of the large fault zone. The great differences in regional stratigraphy across both West and East Penobscot Bays can be explained by this model. The mile-long sliver of Islesboro rocks found outside the area of Fig. 1, 1 mile southwest of the South Penobscot crossroad and 5 miles northeast of the northern tip of the Islesboro block, and the small area of Castine Volcanics south of Marshall Point on northwestern Islesboro indicate right-lateral strike-slip motion on this fault. The fault pattern within the Islesboro block, the many minor structures along the western border of Islesboro and on Ram, Seal, Flat, Warren, and Seven Hundred Acre Islands, and the abrupt and great contrast in metamorphic grade across West Penobscot Bay, also all indicate large right-lateral fault motion.

The Turtle Head fault zone cannot be observed directly on Islesboro as the zone is under water. Breccia zones along the fault trace have been observed in Penobscot Township to the north. Intensely cleaved rocks make up the western shore of Seven Hundred Acre Island, the west shore of Warren Island,

and all of Flat, Seal, and Ram Islands to the north. These localities form a line that transects bedding and minor faults and indicates the approximate trace of the fault zone. The intensity of the cleavage measured by the spacing of slip planes and the amount of mineral reorientation in the schist decreases away from the fault trace, as will be demonstrated by several stops, suggesting that cleavage and faulting formed in response to the same forces.

The fault zone on the east side of Islesboro either is further offshore, or the Islesboro rocks there have not been so highly deformed. Both alternatives probably are true, indicating that more of the fault motion occurred along the western branch of the zone.

Cleavages mapped in the Islesboro block are everywhere very steep to vertical. Only one shallow-dipping cleavage has been observed. This is a relict cleavage and is cut by strong steep cleavage. It can be observed at Stop 7. Close to the trace of the fault zone in West Penobscot Bay, cleavage strikes strictly parallel to the fault trace. Eastward from the fault trace, the strike of the cleavage becomes more divergent, ranging from N.20°W. to N.45°E., but it varies only approximately 25° within each of the small fault blocks.

The strong first cleavage is subparallel to the axial planes of gently plunging folds. Within the strongly cleaved western zone, muscovite and chlorite grew parallel to the trace of the fault zone, oblique to the first axial-plane cleavage in many places, including Stops 2 and 8. This recrystallized muscovite and chlorite may be parallel to the axial planes of steeply plunging folds, which are relatively common close to the western fault trace and are uncommon to the east away from it. I hypothesize that the steep folds and second cleavage formed during fault movement and that the attitude of the cleavage indicates the attitude of the fault zone. Accordingly, the fault zone may be nearly vertical.

The age of movement along the Turtle Head fault zone is known very closely. The fault breaks Castine Volcanics of Late Silurian and Early Devonian age (Brookins, Berdan, and Stewart, 1973) and the South Penobscot pluton (Lower Devonian), which metamorphosed the Castine Volcanics, is intruded by the unbroken Lucerne pluton (Middle Devonian). The Turtle Head fault is thus known to be post-Early and pre-Middle Devonian in age.

The Turtle Head fault zone appears as a low on the aeromagnetic map by Taylor and others (1968) and can be traced southward by this means through the Gulf of Maine until it reaches shore again at Newburyport, Mass., where it separates a Baltic Silurian and Devonian volcanic terrane on the southeast from higher grade metamorphosed rocks of the Merrimack synclinorium to the northwest. Its trace toward the northeast is less certain, but probably it crosses the Canadian border north of St. Stephen, New Brunswick, and extends into New Brunswick at least as far as the St. John River. It may be comparable in scale and rate of movement to the San Andreas fault of California.

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Itinerary

The trip begins at the State of Maine Ferry Service dock at Lincolnville, Maine. Participants must consolidate into as few vehicles as possible so that adequate space will be available on the ferry BOTH ways. Very early arrivals should attempt to see the highly metamorphosed Penobscot Formation typical of the mainland west of Islesboro. Outcrops can be seen north of Lincolnville along Route 1, or west of Lincolnville along Route 173 to Slab City. The outcrops south of Lincolnville are the Megunticook Formation of Osberg (Trip A-4), which reputedly lies beneath the Penobscot Formation. An excellent outcrop of Penobscot Formation containing 3-cm andalusite crystals occurs on the north shore of the mouth of Ducktrap River, about 1 mile north of the ferry landing.

Mileage

0.0 Board ferry at Lincolnville

0.01 Depart ferry at Grindle Point, Islesboro - Drive northeast on paved road.

1.2 Stop sign. Turn left on tarred road.

2.7 Pass access road to airport on right, meeting late arrivals and first class guests.

3.1 Stop sign. Turn left on tarred main road.

5.6 Pass Shell station and Islesboro fire station on right.

6.25 Fork in road, take left road.

7.3 At curve in road, note abandoned paved road on left. Make sharp left U-turn, go 100 yards, and turn right onto dirt road.

8.0 Park in small lot with care. Proceed on foot to shore outcrops north and west of lot.

Stop 1. As the shoreline is reached, massive buff quartzite of the Islesboro Formation intruded by greenstone dikes is visible on the left. DO NOT try to descend to the shoreline here, as this unit will be seen later on the trip. The cove is caused by erosion along a fault that strikes northeast inshore.

Along the right-hand trail from the parking lot, highly sheared gray pelitic rocks can be seen. The cleavage is N.10°E., dipping 80°E. A weakly developed second cleavage is N.10°W., and vertical.

Another 100 meters along the shore trail, a second fault is crossed, and highly cleaved Castine Volcanics are encountered These are purple and green andesitic tuff, breccia, and agglomerate and are typical of substantial thicknesses of

rocks in the Castine section. The cleaved clasts show that strong cleavage may involve only small displacements. No other Castine rocks occur in the Islesboro block. These Castine outcrops are interpreted as a sliver dragged along the fault as much as 13 km from the northeast.

Return to cars, and drive out the same road, recrossing the quartzite terrane.

8.7 Turn left onto main road.

9.7 Outcrop of quartzite on left side of road is the northernmost outcrop of this lithology before it is cut off by faults.

10.1 Park on right side of road by beach. DO NOT TRESPASS on grounds of house on beach side of road. Walk directly from the main road to the tidal part of the beach, and proceed north (right) along the west shoreline of Turtle Head. Take your camera, as this stop is a beauty for future lectures!

Stop 2. A sequence of thin-bedded gray siltstone and sandstone beds has been folded, refolded, and cleaved several times. Gradual northward increase in the intensity of deformation is interpreted to indicate that the fault along the west side of the Islesboro block converges northward on the shoreline. Axial-plane cleavage becomes highly developed, spectacular cleavage refraction is seen across sandy and silty beds, and extensive chemical migration has formed new mineral layering parallel to cleavage. The minerals are chlorite, muscovite, quartz, opaque minerals, and a little albite. Original bedding is increasingly obscured northward, and at the end of the traverse, the extensive development of a second, and even of a third, cleavage makes its recognition very difficult. In view of the limited time for movement along the Turtle Head fault zone, I hypothesize that the sequence of cleavage orientations probably results from rotations of compression axes during fault motion rather than recrystallization during epochs of deformation widely separated in time.

Return to car and drive east on road.

11.4 Water wells in this area pass through 36 meters of glacial till!

11.8 Sarpola residence on left. Turn left down dirt road to end of road.

12.3 End of road, park cars. Proceed past house toward flagpole on point.

Stop 3. This outcrop was cited by Smith, Bastin, and Brown (1907) as best showing the quartzite and quartzite conglomerate (which they called the Battie Quartzite) resting on an impure marble (which they called the Coombs Limestone Member of the Islesboro Formation). Approximately 45 meters of quartzite is exposed, resting on 2 meters of slaty pelite. At least 6 meters of contorted dolomitic marble is separated by 45 meters of beach from the quartzite, but north along the beach at Coombs Point, outcrops of quartzite are within a few meters of outcrops of dolomite.

The section is upright. Top criteria at this stop include sandy beds with siltstone tops and crossbedding. Although many of the clasts in the conglomerate are quartzites, a few gray, black, or red clasts of jasper or chert are present. The axis of the open syncline plunging northeast is sub-parallel to the axis of a larger syncline to the southeast and an anticline 0.6 km to the west.

Return to cars, retrace route to tarred road.

12.8 Tarred road, turn left.

14.5 Rejoin original route at junction.

14.9 Pass Shell station and Islesboro fire station.

16.3 Stop on right side of road by road.

Stop 4. Yeaton D. Randlett property. The outcrop is directly over the bank toward the shore from the back (sea) side of this house. Permission for entry should be sought before going to the outcrop.

The section of green slate and greenstone between the cars and the outcrop contains several impure carbonate beds containing shaly beds. The rocks are tightly folded about nearly vertical axes, and some shearing has occurred. Of special interest is a shaly bed in the carbonate that has been plastically sheared and rotated by a left-lateral couple toward the axial plane. Contortionists can take a beautiful picture here.

Return to cars, and proceed in same direction along the main road.

17.2 Turn right on paved road.

19.2 Turn right at sign pointing toward ferry landing.

20.5 Arrive at ferry landing and park cars. Take only essential field gear and adequate clothing (it's at least 10° colder in a boat). Proceed by chartered small craft to Stop 5, northeast side of central point on Spruce Island, on shore immediately below 'A' frame camp.

Stop 5. Muscovite pegmatite in impure marble. This is the Precambrian pegmatite, noted by Smith, Bastin, and Brown (1907, p. 9), that has been isotopically dated. Two independent K/Ar dates on the muscovite are 594±18 and 599±15 m.y. A Rb/Sr mineral isochron yields 620±18 m.y. (Brookins and Stewart, oral communication, 1971).

The impure marble is more coarsely laminated than is typical at Lime Island. Garnet-bearing schist crops out on the eastern shore of the cove southeast of the dolomite but will not be visited. Similar rocks will be seen at the next stop.

Reboard boats and proceed to next stop, the landing of the Dark Harbor Boat Yard Corporation at Cradle Cove, Seven Hundred Acre Island.

Stop 6. Walk west past the boathouses along the shoreline to examine rocks in the Precambrian section. Individual beds approximately 10 meters thick differ greatly in composition. The presence of thin beds of quartzite and carbonate suggests miogeosynclinal sedimentation. However, feldspathic gneiss, greenstone, and amphibolite indicate some igneous activity in the same area. Medium-grade metamorphism is indicated by large garnets. Equally large andalusites (chiastolite) are visible at low tide in pelitic rocks west of the point beyond the boat yard. Extensive retrograding has almost obliterated most andalusite and some garnets, but characteristic shapes remain.

The age of this Precambrian terrane can only be estimated from scattered Rb/Sr data as approximately 750±100 m.y. (Brookins, written communication, 1972).

Reboard boats and proceed to next stop.

Stop 7. Southern point of Seven Hundred Acre Island. Good photos for camera fans.

The rocks of the Islesboro Formation at this stop are chlorite-grade pelite and grit that contrast strongly in grade and metamorphic style with the nearby Precambrian rocks. Original

bedding is easily seen with graded beds and bedding-cleavage intersections showing that several 10- to 15-meter folds are present and some are overturned. Pebble beds like those in this outcrop can be traced northward across to the northernmost point of the island.

West along the beach 150 meters, two cleavages can be seen in the steeply dipping west-facing strata. Chlorite bands a centimeter apart mark a relict cleavage that strikes N.60°-80°W. and dips northward at about 45°. The relict cleavage is cut by a younger vertical cleavage that strikes N.10°C. The older cleavage is enigmatic as it cannot be related to folds, and elsewhere on the Islesboro block only vertical or steeply dipping cleavage has been observed. The closest flat-lying cleavage known is at Rockport, 11 km southwest of this locality (Osberg, oral communication, 1970), where axial-plane cleavage occurs in over-turned folds at much higher metamorphic grade.

The younger cleavage is parallel to the large western fault zone and is spaced at centimeter intervals. This spacing should be remembered for comparison with that at the next stop, which probably is only about 100 meters stratigraphically from this horizon but approximately 0.6 km closer to the fault zone.

Return to boats. Proceed to point on west side of Philbrick Cove.

Stop 8. Cameras should be taken ashore. The greenish-gray slates of the Islesboro Formation close to the major fault zone show a remarkable development of cleavages and mineral migration. Old quartz veins crossing the slates show ptygmatic folds with right-lateral shear sense, although younger quartz veins are little deformed. Secondary cleavage at small angles to a first cleavage almost obliterates it. The spacing of both cleavage planes is much closer than at the last stop, being 1 to 2 mm. Intensely cleaved rocks like these make up the western shore of the northern point on Seven Hundred Acre Island, the west shore of Warren Island, and all of Flat, Seal, and Ram Islands to the north and mark the inferred trace of the major fault zone.

Return to boats for trip to ferry landing. On the way, note the highly cleaved rocks of the western shores of Seven Hundred Acre and Warren Islands. The Islesboro Formation along this shoreline is a green or gray slate some parts of which contain many ferruginous and calcareous claystone concretions. These resist weathering, and together with the intense cleavage and

differential weathering yield a rasplike outcrop surface that needs to be treated with respect by geologists and stranded seamen.

Trip ends at ferry landing. Line up cars for next ferry to Lincolnville. If time remains, a visit to the museum in the lighthouse at Grindel Point or a walk along the shore would be informative. Walk through the spruce grove past the brick shed northwest of the toll booth. A series of thin to 6-meter thick ferroan dolomite beds, pelites, and greenstones is exposed on the shore. The carbonates are hardly cleaved, with schistosity parallel to bedding. Nearby gray pelites and psammites are cleaved, and greenstone dikes, though broken into blocks, are hardly sheared internally. Although this area is only a few hundred yards from the projected strike of the highly cleaved rocks on Warren Island, it has not been strongly affected. Possibly the Grindel Point-Islesboro Harbor sub-block moved southwestward into the fault zone, and the cleaved rocks were scraped off shortly before fault motion ceased.

Drive carefully up Route 1 to Stockton Springs, 1a to Bangor, and 95 to Orono.

IGNEOUS PETROLOGY OF SOME PLUTONS IN THE NORTHERN PART OF THE PENOBSCOT BAY AREA

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Introduction

This field trip is designed to examine variations among some of the plutons in the Penobscot Bay area. The plutons, in general, form topographic highs on the mainland and dominate the geology of the major islands: Vinalhaven, Isle Au Haut, Deer Isle, Swans Island, and Mount Desert. Field observations, combined with appropriate laboratory studies, can provide estimates of depth of crystallization, variations of magmas in time and space, and, perhaps, absolute ages of the events within the intruded terrane.

These plutons have intruded a zone some 50 km (31 miles) wide (see Fig. 1; also Fig. 1, trip B-7) of fault bounded blocks that contain disparate lithologies including banded gneisses, marine sedimentary rocks, and volcanic rocks. (See trip B-7.)

The area may represent a plate boundary in Paleozoic time (Bird & Dewey 1970; Naylor 1971; Wilson, 1966). Magma composition may be related temporally to this feature, but there seems to be no valid documentation of a relationship between geologic setting and the composition of the intruding magmas.

In this discussion the new International Union of Geological Sciences (IUGS) recommendations on nomenclature are being followed and the term granite includes all rocks containing more than 20% quartz and having an alkali feldspar to plagioclase ratio greater than 0.5. Modes of some plutons are given in Table 1 and are plotted in Figure 2.

I would like to acknowledge the help of R. M. Hazen, D. M. Miller, and R. L. Trithart in the field studies, and of N. L. Hickling, E.G. Williams, and C. S. Zen in the laboratory studies. P. H. Osberg, D. W. Rankin, and D. B. Stewart gave generously of their time to help make this a better guidebook article.

Stratigraphy and Geochronology

The plutons of the area can be grouped into four types (Fig. 1), although the divisions of this classification scheme are based on both age and petrography and are somewhat arbitrary. Two plutons, the South Penobscot and the Mount Desert, contain rims of dioritic to granodioritic material which contains inclusions of mafic rocks and is intruded by later biotite and (or) hornblende granites. These two plutons are also sheared and faulted (Chapman, 1970; see trip B-7) so that significant tectonic activity post-dated their intrusion. The core materials are

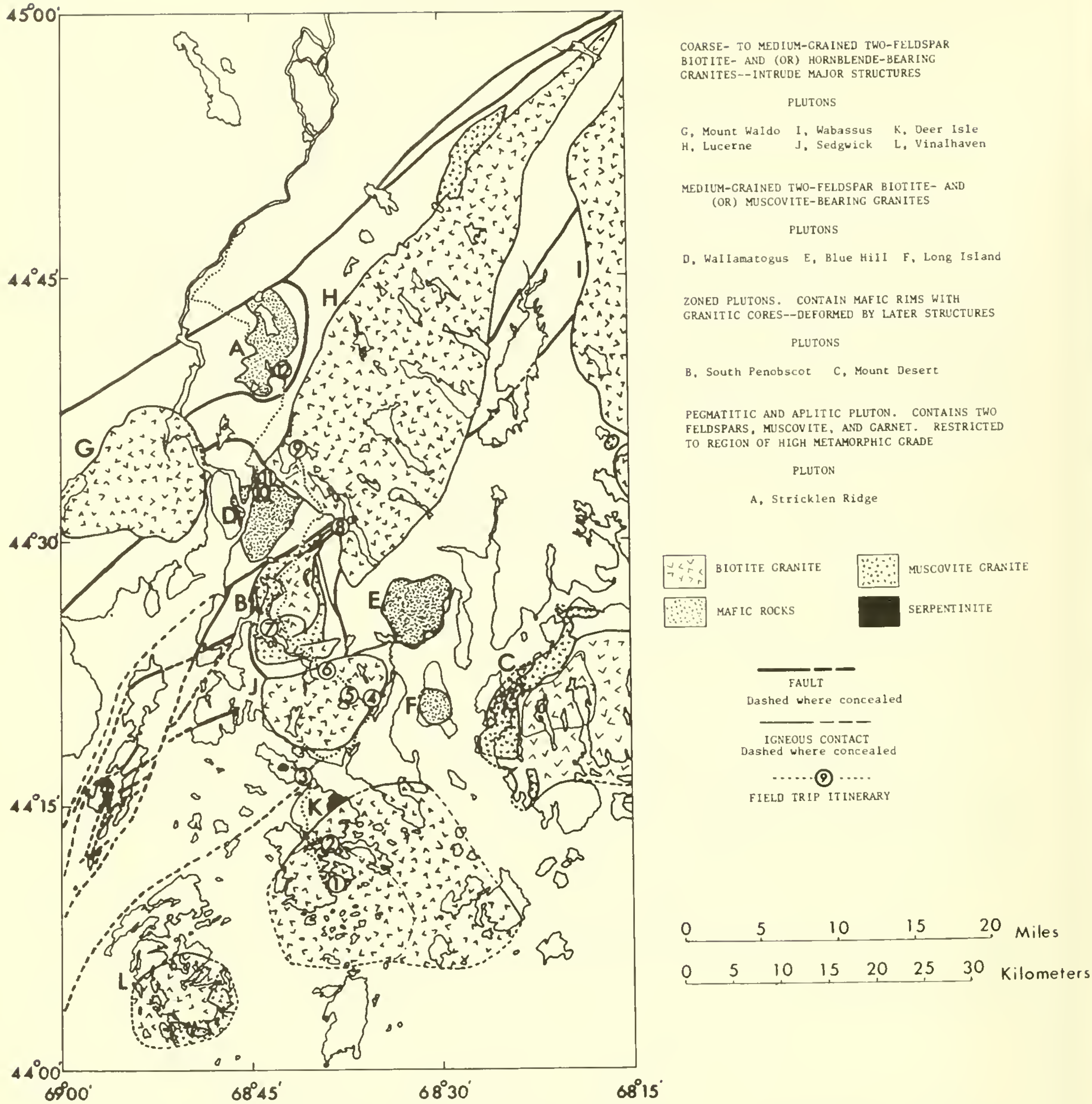

Figure Captions

Figure 1. Sketch map showing locations of plutons in the Penobscot Bay area, Maine. For geology of country rocks see Figure 1, trip B-7.

Table 1.--Modes of some granitic rocks in the Penobscot Bay region

	1	2	3	4	5	6	7	8	9	10	11	12	13	14
Quartz	28.13	28.02	24.8	31.33	25.43	8.3	25.1	31.17	33.09	30.26	28.35	32.95	28.84	23.68
Alkali feldspar	30.83	46.01	27.2	38.50	36.68	6.8	25.4	7.75	38.04	36.53	35.68	25.60	47.95	24.27
Plagioclase	35.06	24.44	39.4	24.15	28.20	65.3	41.4	50.42	19.89	26.66	27.60	28.59	14.33	41.67
Muscovite							1.4					5.74	5.97	1.42
Biotite			5.2			19.2	4.4	10.08				6.95	2.90	8.79
Chlorite			0.7											
Mafics[1/]	5.99	1.55		6.04	9.70				6.16	6.55	8.37			
Amphibole			0.4					0.80						
Opaques			1.2			0.3		0.08						

1. Stonington granite, Buckmaster Neck (Stop 1)
2. Aplite, Buckmaster Neck (Stop 1)
3. Oak Point Granite (of local usage) Mountainville (Stop 2)
4. Sedgwick granite (Stop 4)
5. Sedgwick granite (Stop 6)
6. South Penobscot pluton, matrix at Johnson Pt. (Stop 7)
7. South Penobscot granite, north of Johnson Pt. (Oliphant, 1969)
8. South Penobscot granite, inclusion in Lucerne Granite (of local usage) (Stop 8)
9. Lucerne Granite (of local usage) (Stop 9)
10. Leucocratic dike (Stop 9)
11. Lucerne Granite (average of 17 modes)
12. Wallamatogus granite, interior facies
13. Wallamatogus granite, rim facies (Stop 10)
14. Stricklen Ridge pluton, aplitic facies (Stop 12)

1/ Mafics includes all dark minerals.

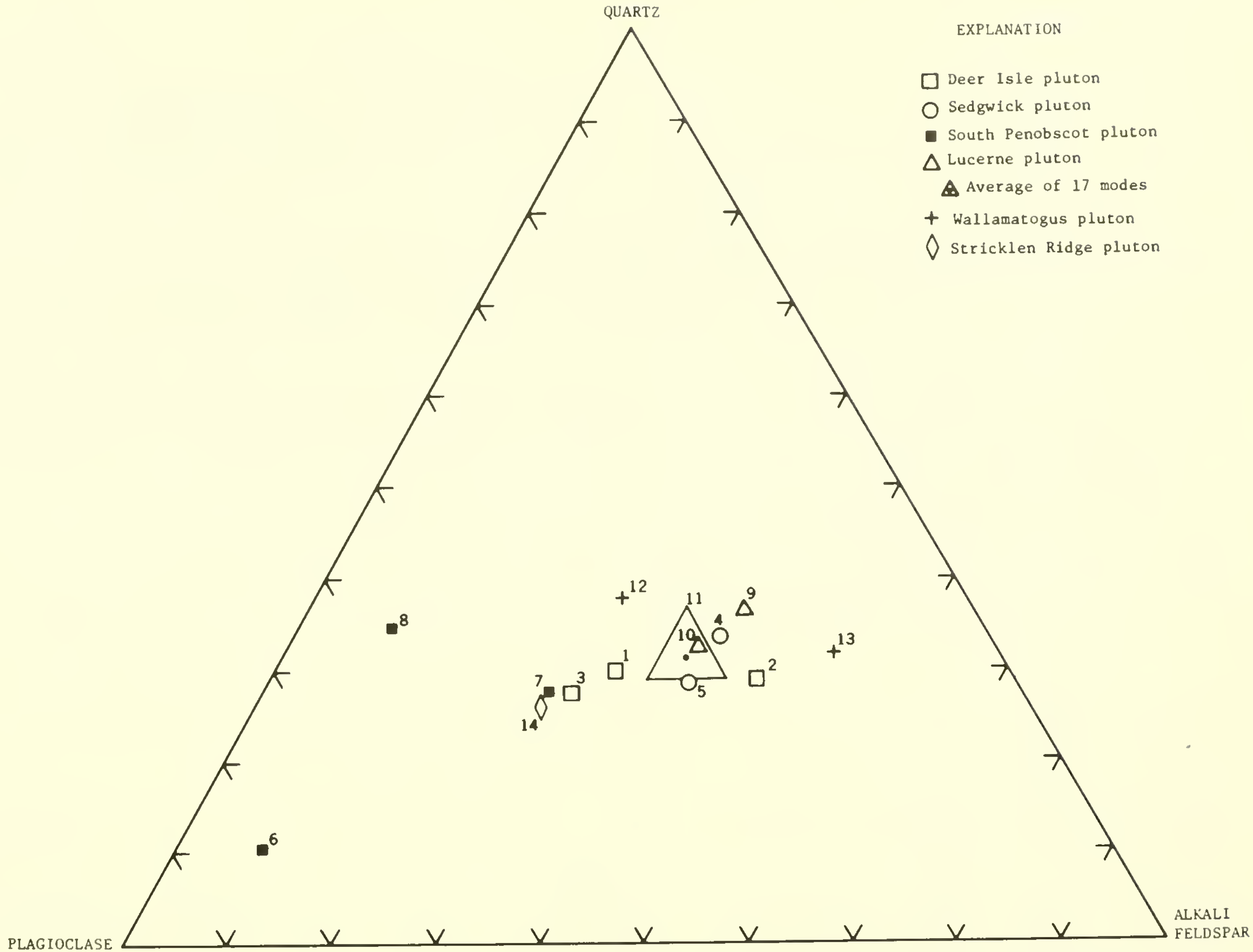

Figure 2. Quartz-plagioclase-alkali feldspar modal plots for plutons in Penobscot Bay area, Maine. Numbers correspond to columns in Table 1.

biotite and (or) hornblende granites which intrude the earlier rim material.

The Wallamatogus, Blue Hill, and Long Island plutons contain muscovite granite and tourmaline-bearing zones. The metamorphic aureole of the Wallamatogus pluton is more extensive than that of the Lucerne pluton. Although some minor shearing has been observed in the Wallamatogus pluton, these plutons are faulted to a lesser extent than the South Penobscot and Mount Desert plutons.

Biotite granites characterize the Mount Waldo, Lucerne, Wabassus, Sedgwick, Deer Isle, and Vinalhaven plutons. The granites within these plutons are generally medium to coarse grained and tend to have porphyritic facies with alkali feldspar phenocrysts. The Lucerne (Cavalero, 1965) and Vinalhaven (Smith, 1896) plutons have earlier mafic intrusives associated with them, but not the actual rim of diorite and granodiorite characteristic of the South Penobscot and Mount Desert plutons. The Deer Isle pluton has a hornblende-bearing rapakivi partial outer rim (Oak Point Granite, Stewart, 1956) which grades into the Stonington granite of the core. Unlike the muscovite granites, the biotite granites all seem to have narrow thermal aureoles as compared to their diameters.

The Stricklen Ridge pluton is quite different from the others in the region. It pervades, and is intimately mixed with inclusions of, the Passagassawakeag gneiss and the Copeland schist. It has a highly variegated texture ranging from aplite to pegmatite. Garnet and muscovite are ubiquitous, but biotite is less common. The pluton and its apophysal dikes are truncated by faults and all are confined to the fault block containing the Passagassawakeag gneiss (see Field Trip B-7).

Geologic evidence clearly makes the Lucerne pluton younger than the Wallamatogus and South Penobscot plutons. Indirect evidence also implies that the Mount Waldo pluton is younger than the Wallamatogus pluton. Other than these limits, the relative ages of the plutons cannot be established by geologic criteria. Brookins has compiled ages from the plutons derived from Rb-Sr and K-Ar methods and Table 2 results largely from his efforts. He has also provided a brief appendix for this excursion.

All of the plutons discussed herein, with the exception of the Stricklen Ridge pluton, are 375 ± 25 m.y. old and are Devonian. The isotopic age resolution is not good, but it appears that in the Penobscot Bay area, granitic intrusions are continuous rather than episodic. As can be seen by Table 2, the isotopic ages are not sufficient at present to resolve the relative time scale of these intrusions.

Preliminary work on the Stricklen Ridge pluton by Brookins (oral commun., 1974) indicates at least an Ordovician age, and the youngest migmatites in the Passagassawakeag gneiss contain zircons in which Zartman (oral commun., 1974) obtained preliminary Pb^{206}/Pb^{207} age of 430 ± 10 m.y.

Table 2.--Isotopic age determinations for some plutons in the vicinity of Penobscot Bay, Maine

Pluton	Age	Method	Investigator
Lucerne	356	K-Ar, biotite	Faul *et al.*, 1963
	402±13	Rb-Sr WR	Brookins, written commun. 1974
Mount Waldo	390±10	Rb-Sr WR	Brookins, in press
Wabassus	370	K-Ar, biotite	Faul *et al.*, 1963
Sedgwick	413±15	K-Ar, biotite	Brookins, in press
	395±15	Rb-Sr WR	Brookins, in press
Deer Isle	357±1	Rb-Sr WR	Brookins & Spooner, 1970
	355	K-Ar, biotite	Faul *et al.*, 1963
Vinalhaven	361±7	Rb-Sr WR	Brookins, in press
	399	K-Ar, biotite	Faul *et al.*, 1963
Wallamatogus	375±20	Rb-Sr WR	Brookins, written commun.
	384	K-Ar, muscovite	Faul *et al.*, 1963
Blue Hill	387	K-Ar, biotite	Faul *et al.*, 1963
	382±10	Rb-Sr WR	Brookins, in press
Mount Desert	394	K-Ar, biotite	Faul *et al.*, 1963
	429±13	Rb-Sr WR	Metzger & Bickford, 1972
South Penobscot	360±15	K-Ar, biotite	Brookins, in press
	403±15	Rb-Sr WR	Brookins, in press
Migmatite	430±10	Pb^{206}/Pb^{207}, zircon	Zartman, oral commun., 1974

Each pluton has its own particular characteristics as to modal variations, foliation, fabric, inclusion content, aplite content, aplite distribution, and alteration zones. Red coloration is common in the Vinalhaven, Deer Isle, and Mount Desert plutons, whereas the Stricklen Ridge, Wallamatogus, and South Penobscot are white. Red and salmon-colored areas are common in all of the other plutons.

Aplite or leucocratic late-stage dikes are very common in the Vinalhaven and Deer Isle plutons and relatively uncommon in the Wallamatogus and Lucerne plutons. Magnetite is a common accessory in all of the biotite granites, except for the Lucerne pluton whose contacts can be mapped under the glacial cover by its characteristically low magnetic signature (Kane _et_ _al_., 1971).

Relation of Plutons to Structural Setting

The plutonic varieties are not particularly related to the regional structures. In this area, the plutons are all located southeast of the "Norumbega" fault (see trip B-7), but outside this area, to the southwest and northeast, plutons occur on both sides of the "Norumbega" fault.

The South Penobscot and Mount Desert plutons seem to contain more mafic inclusions than the other plutons, and these inclusions could represent the southwestern extremity of Chapman's (1962) Bays-of-Maine complex.

Muscovite granites are notably less abundant in the Penobscot Bay area than they are in Western Maine or New Hampshire where binary granites are more common (Billings, 1956). There are no significant petrographic differences between plutons intruded on opposite sides of identified faults.

The physical appearance of a pluton is a complex function of its composition, rate of cooling, and interaction with its surroundings. As we shall see, the Lucerne pluton contains few aplites, is generally white and exceedingly coarse grained, and contains very little magnetite. The Deer Isle pluton, in contrast, contains many aplites, is reddened and medium grained and contains magnetite as a common accessory. What do these observations mean? Was the Lucerne intruded at a deeper erosional level, did it contain more or less volatile constituents, or did it cool more slowly than the Deer Isle?

Sweeney (1973) has considered the shapes of these plutons on geophysical evidence and his ideas are presented in an appendix.

The general lack of significant volumes of granodiorite or tonalite makes it unlikely that much oceanic crust was partially melted to yield these magmas. They appear to represent fusion of continental crust. The most appropriate present analogs along modern plate boundaries might be either the volcanic centers along the San Andreas fault system of California or those in the Aegean Sea.

The trip will examine the Deer Isle pluton (Stops 1 & 2), the Sedgwick pluton (Stops 4 & 6), the South Penobscot pluton (Stops 7 & 8), the Lucerne pluton (Stops 8, 9, & 11), the Wallamatogus pluton (Stops 10 & 11), and the Stricklen Ridge pluton (Stop 12).

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Itinerary

Mileage

0 Field trip begins at Perini Corp. quarry on Buckmaster Neck. Quarry. Buckmaster Neck, Oceanville, Stonington.

Stop 1. This property belongs to Perini Corp. who have graciously granted us permission to visit it. BE CAREFUL— the blocks in this quarry are in a metastable condition. They move easily; be sure you are not under one when it moves. Blocks on the west side of the quarry have moved 5-20 cm. since quarrying activity stopped. This quarry is located in the Stonington granite within the Deer Isle pluton. The quarrymen exploited the exfoliation joints and most recently used flame spallation method to cut the blocks. Note the boom for transferring blocks from the working face to the loading dock where they are weighed. For a variety of economic reasons, natural building stones have been supplanted by concrete and structural steel.

The grain size is highly variable with the alkali feldspar giving the granite a porphyritic quality. Pegmatitic pods contain feldspars as large as 2-4 cm. Sparse rapakivi-textured feldspars are present. The salmon perthite combined with the dark biotite, white plagioclase, and gray quartz make this an interesting and attractive building stone.

Late-stage aplite dikes are common. Most have an accumulation of mafic material on the footwall such as Quinn (1943) described for the Westerly, R. I., granites. However, one

aplite on the upper level of the quarry has mafic accumulations on both the hanging wall and the footwall. Footwall segregations alone are due to gravitational settling, but segregations on both walls must be due to flow segregation. Inclusions are rare and highly variable. On the east wall, one can see how the granite near the surface is reddened along the joints; the color is due to disseminated flakes of hematite within the feldspar.

0.9 Turn right on Route 15 north.

1.8 Cross Inner Harbor Bridge into town of Deer Isle.

4.4 Turn right toward Sunshine.

5.8 Turn right toward Sunshine.

7.0 Park on left, walk ahead across Greenlaw Cove bridge to roadcut.

Stop 2. GREENLAW COVE. This road cut gives an excellent exposure of the Oak Point Granite (of local usage), the eastern and outer member of the Deer Isle pluton. Rapakivi texture is especially well developed here. Stewart (1956) and Stewart and Roseboom (1962) have discussed the magmatic origin of this texture using this granite as the basic observation. Note the large number of aplite dikes and the bright red-orange color of the alkali feldspars. Note the abundance of inclusions and the development of rapakivi porphyroblasts within the inclusions. Alteration is well developed along joints. The radial fracture pattern well developed here results from blasting for the roadcut.

Return to outcrops north of parking area. The intensely altered zone is now quartz and maximum microcline as opposed to the quartz, plagioclase, biotite, and orthoclase (and (or) intermediate microcline) of the original rock. The feldspars are homogenized, not perthitic; quartz veins are common, and the quartz is not milky. Aplites are not easily observed. These alteration zones are strongly magnetic.

Turn around and return to Route 15.

9.6 Turn right on Route 15.

10.1 Bear right toward Sedgwick.

12.4 Causeway at Deer Isle and Little Deer Isle.

Stop 3. The Causeway is made up of serpentized dunite quarried from the highest hill visible on nearby Little Deer Isle. The serpentite body has no exposed contacts,

and it is uncertain whether it represents an ultramafic intrusion or a block of ultramafic material faulted into its present position.

Continue north on Route 15.

14.3 Deer Isle information center; continue right on Route 15.

15.0 Deer Isle bridge across Eggemoggin Reach.

16.5 Continue straight on Route 175 south toward Blue Hill.

20.0 Turn left on Route 172 north toward Ellsworth.

23.6 Bear left on old road and park.

Stop 4. This roadcut is in the southeastern part of the Sedgwick pluton. The elliptical outcrop pattern results from this nearly cylindrical mass plunging about 25° SE. In contrast to the Deer Isle pluton, notice the pale color and lack of aplitic material. A small red pegmatite seam containing muscovite and tourmaline is present. This fracture is interpreted to be a post-magmatic feature involving recrystallization of the granite in the presence of gas. The presence of muscovite here, and not in the granite proper, is attributed to the incongruent solubility of feldspar in the gas phase. Contrast this with the Wallamatogus granite (Stop 10).

23.7 Turn right on Route 172 south.

24.4 Sedgwick Grange. Turn right on Ridge Road.

27.2 Secondary Road on south side. Park on Ridge Road.

Stop 5. This road, built a decade ago by the operators of the Black Hawk mine in Blue Hill to supply access to a drilling program, is paved with sulfide ore from the mine. This brief stop is to provide you with a look at the mineralogy and assemblages within this copper-zinc-lead deposit.

Continue northwest on Ridge Road.

28.5 Intersection Route 176. Park in equipment park on left.

Stop 6. Northern contact of the Sedgwick pluton. In contrast to Stop 4, the Sedgwick granite is here coarser, and filled with inclusions. This would be the "footwall" of the plunging cylinder. Xenoliths are very common here. The contacts of the biotite granite plutons are abrupt and generally show only a few apophysal dikes. In contrast,

the muscovite-bearing granites send out many apophysal dikes and tend to metasomatize the surrounding schists.

Turn left and continue on Route 176 west.

29.1 Turn right on Route 176 west towards Penobscot.

31.6 Go straight on Route 175 north towards Orland.

33.4 Cross brook and turn left on paved road.

34.3 Bear left past gray house marked "Webb."

34.5 Park in field. Outcrops are 200 m (660 feet) north along shoreline.

Stop 7. South Penobscot pluton, western border phase. This pluton has a rim of diorite containing many mafic inclusions. The inclusions range from gabbro to pyroxenite. Diabase intrudes the Castine Volcanics and in turn is intruded by the South Penobscot pluton which may be cosanguineous with the Castine Volcanics. The dioritic matrix is dominated by plagioclase and biotite and could have evolved from a mafic magma. Alternatively, a granitic magma could be contaminated by inclusions of gabbroic material, through which the magma has moved. This diorite grades into a granodiorite, which is intruded by a late central stock of biotite granite.

Return to Route 175.

35.6 Turn left on Route 175.

38.7 Turn right on Route 199 toward North Penobscot.

43.8 North Penobscot. Turn right on Route 15 south.

43.9 Turn left on paved road "Balsam Cove Campground."

44.5 Turn right on gravel road.

45.1 Turn right on gravel road marked "Lord."

45.8 Turn around area. Parking here is poor.

Stop 8. Dike of Lucerne Granite (of local usage) intruding a faulted contact between Penobscot Formation and the South Penobscot pluton. Outcrops in woods to east of blueberry field are Penobscot Formation. Outcrop in road is a dike of Lucerne Granite containing inclusions of the South Penobscot pluton, granodiorite phase. Note the very coarse grain size of the Lucerne Granite, even in the

contact areas. This indicates either very slow cooling, or a low viscosity magma which permits rapid diffusion. The inclusions of South Penobscot have very little alkali feldspar (see Table 1). Contrast the inclusions of Penobscot Formation within the Lucerne with the Penobscot Formation in Stop 11.

Return towards North Penobscot.

46.5 Turn left on gravel road.

47.0 Turn right at chicken factory on Back Ridge road.

50.3 Cross U.S. Route 1, East Orland, continue straight ahead to Craig Brook National Fish Hatchery.

51.8 Craig Brook National Fish Hatchery. Continue straight to nature trail.

52.1 Park cars in parking area provided at nature trail. Follow Yellow Dot trail to the right.

Stop 9. Nature trail at Craig Brook U.S. National Fish Hatchery.

500'. Notice gravel outwash is predominately feldspar.

810'. Signpost 5. Friable Lucerne Granite. Gorge dissects massive granite. Sheeting is on the dimensions on 1-4 cm which is the size of the feldspars within the Lucerne. Friable Lucerne Granite is the chief source of road metal in the region between Bucksport and Ellsworth.

870'. As you cross the stream, note that the finer grained leucocratic dike is more resistant than the coarser Lucerne Granite.

940'. Signpost 6. Note how leucocratic dike (N.60°W., vertical) acts as a dam in the stream bed. Note also concentration of dark minerals along edge of dike. An inclusion of coarse Lucerne Granite is within the dike.

Those who wish to return to the parking area may do so via the Blue Dot trail.

1110'. Signpost 8. Leucocratic dike in stream bed. These dikes have tourmaline as the dominant mafic phase.

1200'. Cross stream bed. Note boulder of salmon-colored Lucerne Granite.

1600'. Large block of Lucerne Granite beginning to work loose. Many low ridges in the outcrop area of Lucerne are rubble piles of similar boulders. They imply bedrock nearby, even if they themselves are not in place.

2000'. Descent into the gorge on south (right) side of trail. Here is a typical "pavement" outcrop of Lucerne Granite.

Note 1) Large size of feldspars.
2) Irregular patches of quartz.
3) Rapakivi texture of feldspar.
4) Differential weathering of feldspar.
5) Golden luster of weathered biotites.
6) Foliation and lineation (N.20°E.) caused by alignment of feldspar.

Continue 300' downstream past fallen trees and large boulders. Undercut bank showing differential resistance to weathering of the Lucerne. The resistant area consists of red feldspar and is silicified.

Climb up north bank (right as you face downstream) and rejoin Yellow Dot trail to the east.

3100'. "Ledges" of Lucerne Granite. (When asking Maine natives for outcrops, one always asks for ledge, not bedrock.)

3200'. Signpost 13. Residual boulders of Lucerne.

3450'. Signpost 15. Turn right on Woods road.

4900'. Return to parking lot.

Return to U.S. 1.

53.7 Bear right.

53.8 Turn right on U.S. 1.

54.5 Turn left on paved road "Speed Limit 35."

55.1 Turn right on Route 15.

55.2 Turn left at "Live Bait."

56.9 Bear right.

57.5 Ledge of Wallamatogus interior muscovite-bearing facies on left side of road.

58.1 Park on right side of road.

Stop 10. Rim facies of Wallamatogus granite. The porphyritic phase of the Wallamatogus is sheared here, indicating some post-intrusion strain. Note muscovite crystals which appear to be interlocking and contemporary with the rest of the groundmass. Primary muscovite indicates a significantly more hydrous and peraluminous magma than that which crystallized to the biotite granites.

58.4 Cranes Corners. Turn right on Route 175 north.

59.1 Turn right onto farm belonging to Gilman Harriman. Drive up the driveway and park behind the house. Follow the track 1400'. At this point a large white pine lies S.42°E. Bear S.21°E. 500' to outcrop. Stay in single file and protect blueberries.

Stop 11. Metasomatized Penobscot Formation. This outcrop is gneissic Penobscot infiltrated by myriad fine-grained irregular dikes. Two dikes of Lucerne Granite transect the complex. In places some of the earlier (Wallamatogus?) dikes appear to coalesce with the Lucerne, but most are clearly older. Fragments of Wallamatogus may be found in the Lucerne 2000 m (6600 feet) east of this locality, establishing the Lucerne as the younger of the two granites. The Wallamatogus magma infiltrated the host rock and added potassium to form alkali feldspar within the Penobscot Formation.

59.2 Turn right onto Route 175 north.

60.8 Cross U.S. 1 onto Route 46 north.

61.1 Turn right onto Route 46 north.

65.8 Turn left toward East Bucksport Church.

66.0 Turn right. Drive through area of large gravel pits. These "ice contact" deposits are at about 200' above sea level and represent glacial outwash channeled here to the southwest because of the resistant hills made up of Lucerne Granite to the east. The major lakes in the area all trend to the southeast and are parallel to the observed glacial striae.

67.1 Turn right.

68.4 Notice hills to the east which are outcrops of the Lucerne pluton.

68.7 Park along road. Note outcrops on left.

Stop 12. Stricklen Ridge pluton. This pluton is not well exposed except in the woods. This is the only exposure we shall see and it is not typical. It contains much more biotite than the usual outcrop of this granite in which muscovite and garnet are the common accessories. ALTHOUGH BETTER OUTCROPS MAY BE SEEN IN THE HORSE PASTURE WEST OF THE ROAD, YOU ARE REQUESTED NOT TO GO ONTO THAT PROPERTY IN A LARGE GROUP. The Stricklen Ridge pluton is intimately mixed with the pendants and inclusions of the host rocks, which are either Passagassawakeag gneiss or, as is the situation here, the Copeland Formation. In the pasture west of the road, fragments of an amphibolitic layer can be seen which more or less outlines the anticlinal structure at the northeast end of the horst in which the Passagassawakeag gneiss, Copeland schist, and Stricklen Ridge granite are found.

69.1 Bear left.

72.2 Turn right on paved road.

73.7 Continue straight west.

75.0 Mylonite zone in Passagassawakeag gneiss.

75.6 Continue straight on gravel road.

75.8 Cross major fault zone.

76.1 Cross Johnson Mill road.

77.6 Turn right onto Route 15 north.

78.5 R.R. Crossing.

83.1 Intersection Route 1A. End of trip. To reach Orono follow signs to I-95 North.

IGNEOUS PETROLOGY OF SOME PLUTONS IN THE
NORTHERN PART OF THE PENOBSCOT BAY AREA

Appendix A

Subsurface Density Distributions East and Northeast of Penobscot Bay, South-Central Maine

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Introduction

Gravimetric studies over granites and adjacent supracrustal rocks to the east and northeast of Penobscot Bay in south-central Maine* has defined the configuration of the granitic intrusives, contributed to our knowledge of their mode and time of emplacement and improved our knowledge of exposed and unexposed lithologies.

All felsic intrusives are associated with well defined gravity lows (Figs. 1 and 2) which has resulted in the determination of the following gravimetrically derived properties (granites are discussed in order of their occurrence moving south to north):

The Deer Isle pluton (Stonington and Oak Point granites) are up to four km thick with vertical near surface contacts and a well defined deep zone offset to the northern half of the model (Fig. 3). The model together with the pattern of isoanomaly contours over the body (Fig. 1) suggests that the two granites remain in contact in the subsurface.

The Sedgwick pluton center of gravity occurs under the southern half of its exposure (Fig. 1) where the association of the maximum anomaly gradient with the wallrock contacts indicates their near vertical nature. The northern half of the body has inward dipping contacts (Fig. 3) and, overall, the intrusive is up to five km thick.

The anomaly pattern over the South Penobscot pluton (Fig. 1) indicates a symmetrical mass distribution with respect to its exposure (at least in a north-south direction) with steep but inward dipping contacts. Model results (Fig. 3) bear this out. However, the surrounding geology is complex and subsurface density distributions adjacent to this pluton are, therefore, uncertain.

The gravity pattern over the Wallamatogus pluton (Fig. 2) cannot be separated from that due to the adjacent Lucerne pluton, therefore, the two plutons are modeled together (Fig. 4). The intervening septum of dense metasediments (ρ = 2.79 g/cm^3) has minimum effect on the anomaly pattern between the two plutons (Fig. 2) which indicates that the metasediments

*Gravity data over the Deer Isle pluton and the Sedgwick and South Penobscot plutons were obtained and analyzed by Abbey (1972).

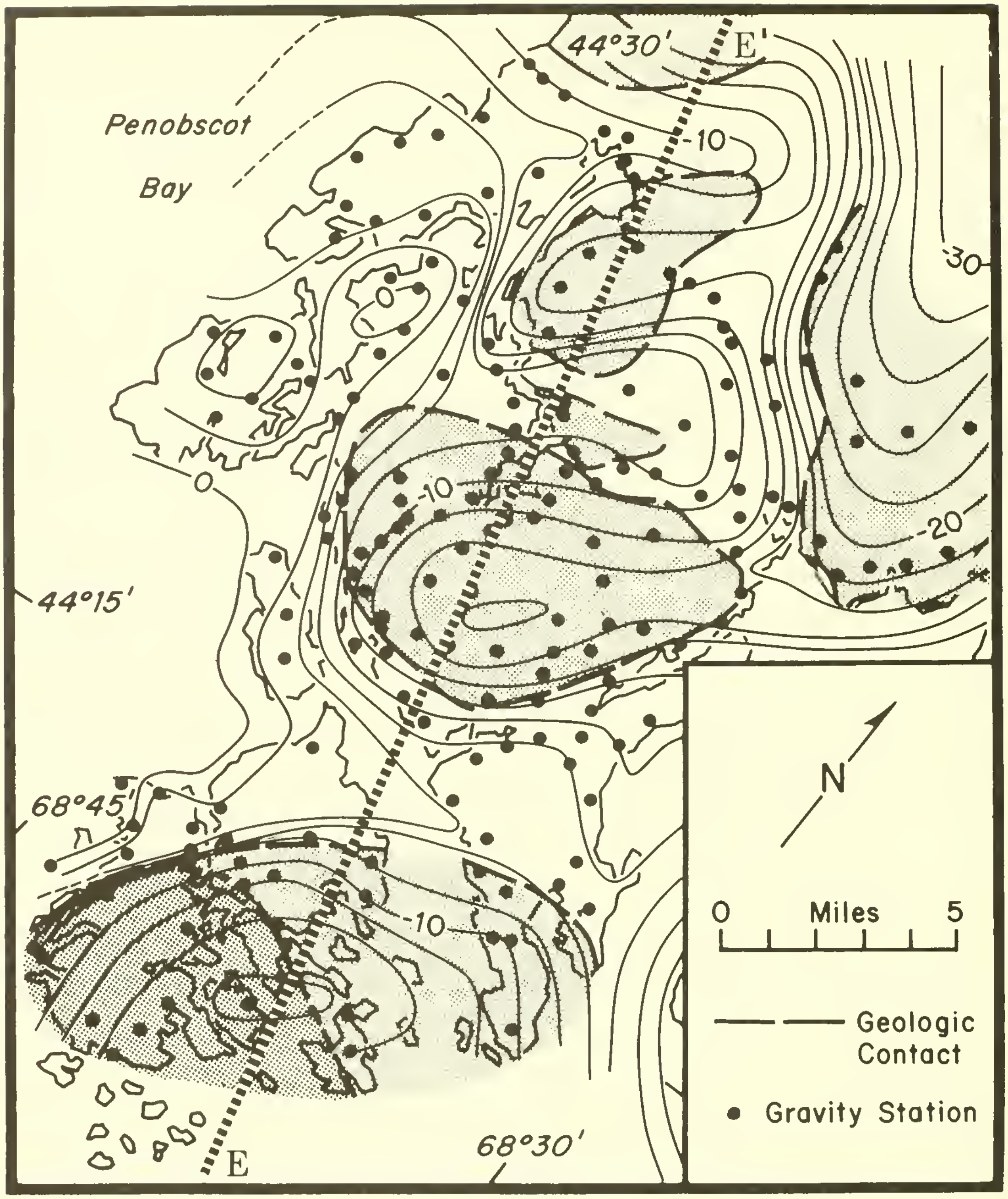

Figure 1. Bouguer anomaly map (two mgal contour interval) over the Deer Isle granites and, to the north, the Sedgwick and South Penobscot plutons. Redrawn from Abbey (1972).

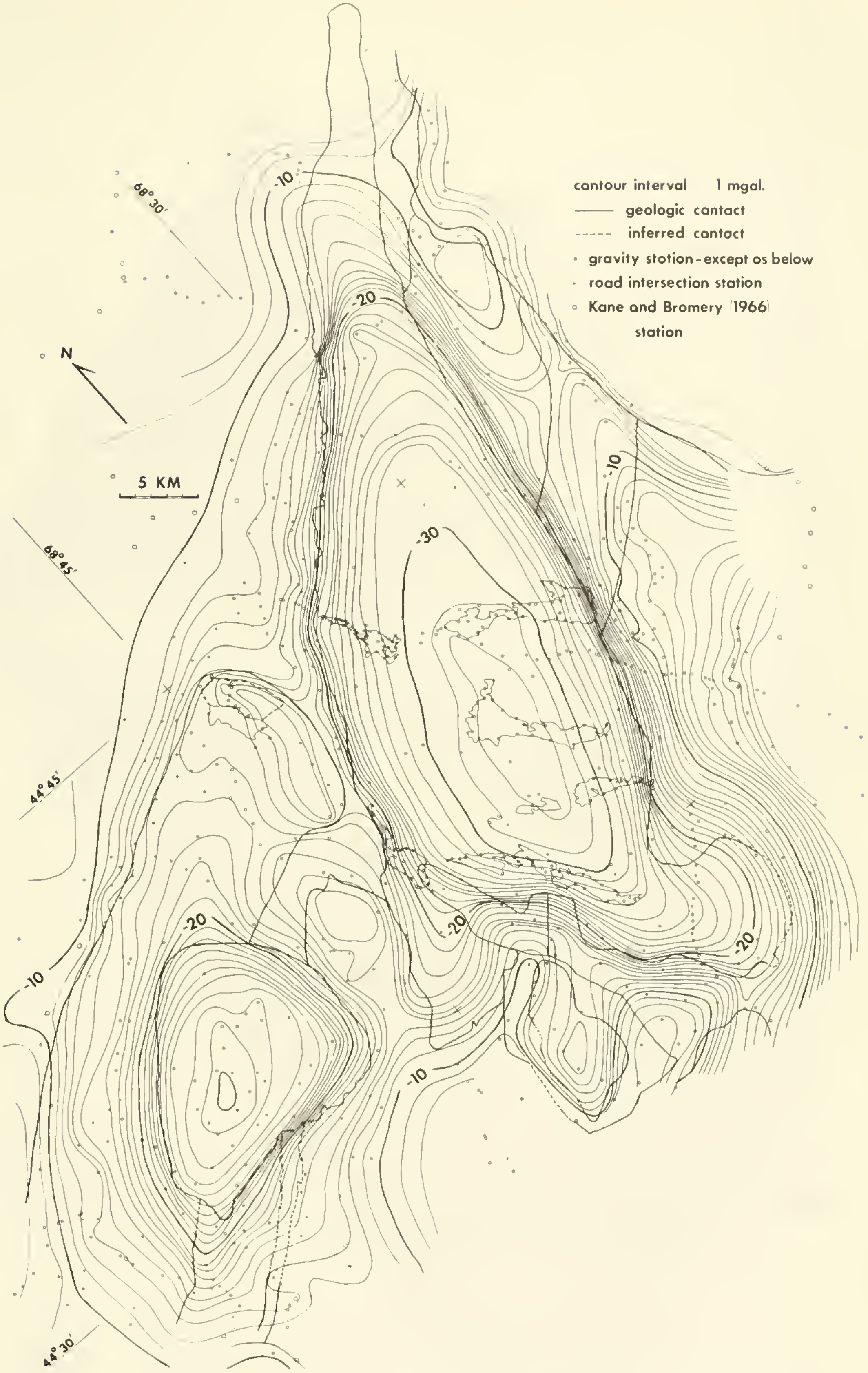

Figure 2. Bouguer anomaly map over the Wallamatogus, Lucerne and Stricklen Ridge plutons.

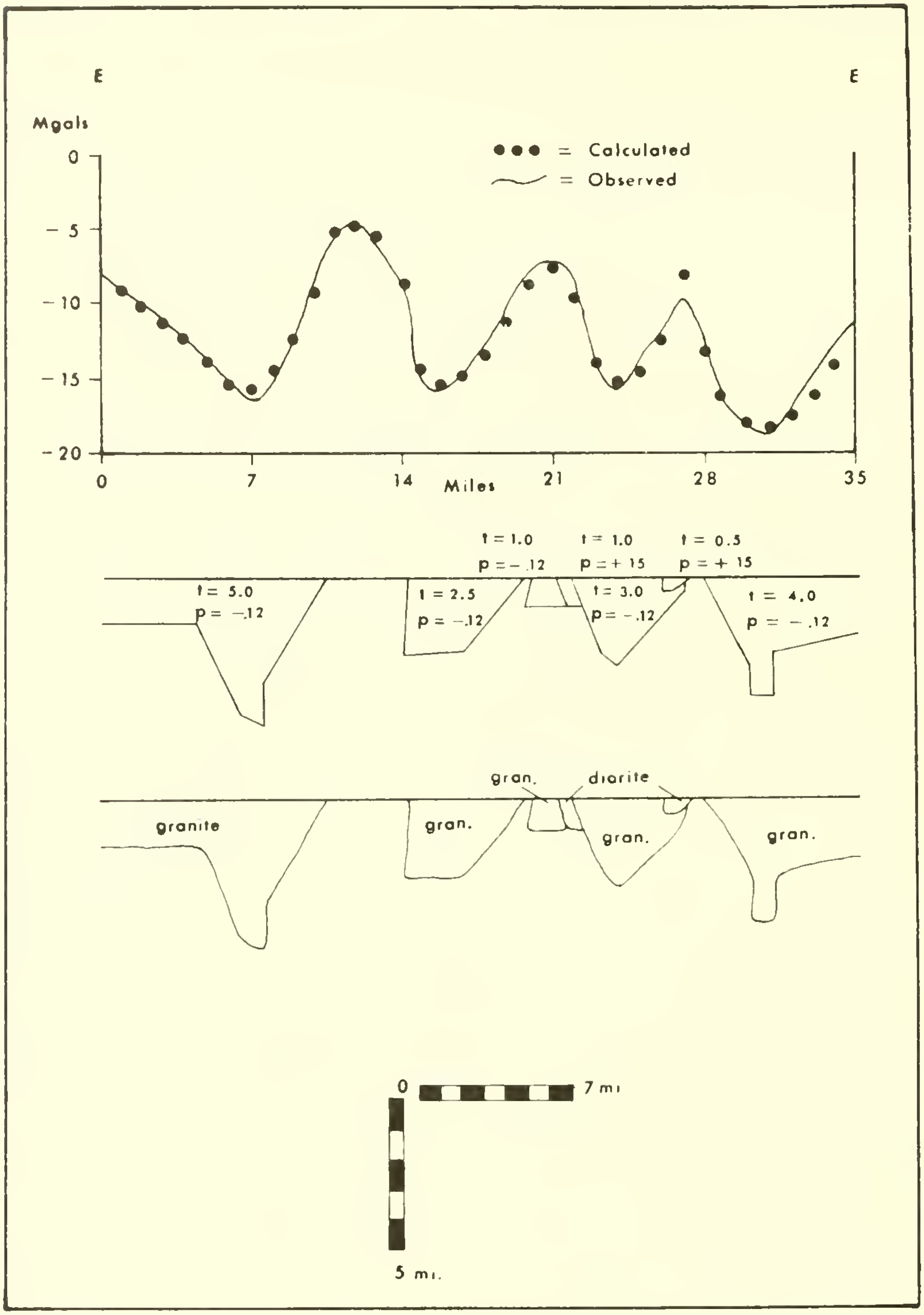

Figure 3. Two-dimensional gravity model along E-E' (see Fig. 1). Reproduced from Abbey (1972).

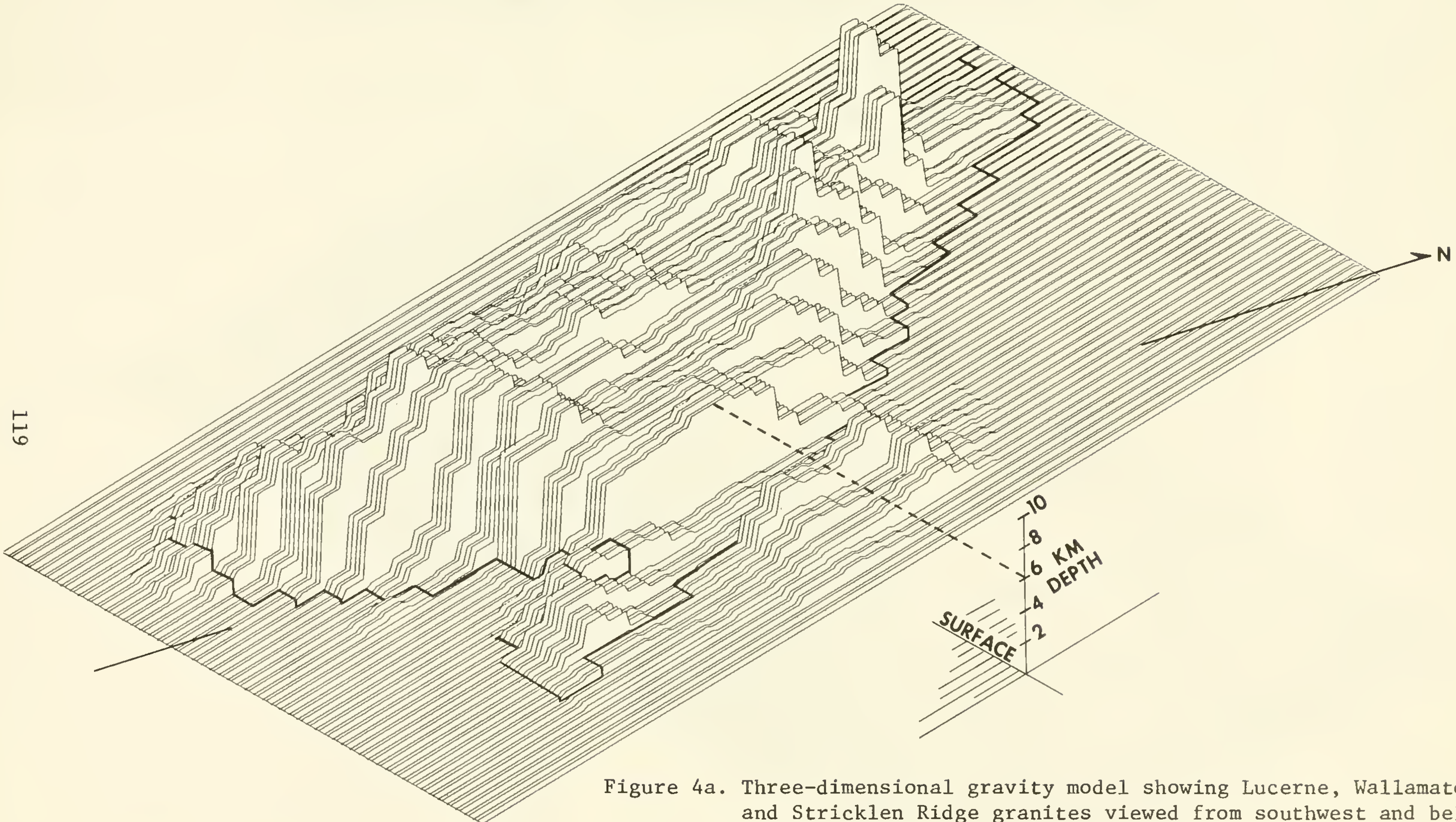

Figure 4a. Three-dimensional gravity model showing Lucerne, Wallamatogus and Stricklen Ridge granites viewed from southwest and below surface. Depth scale west of body. $\sigma = -0.12$ g/cm^3. Dark lines are surficial granite contacts. No vertical exaggeration.

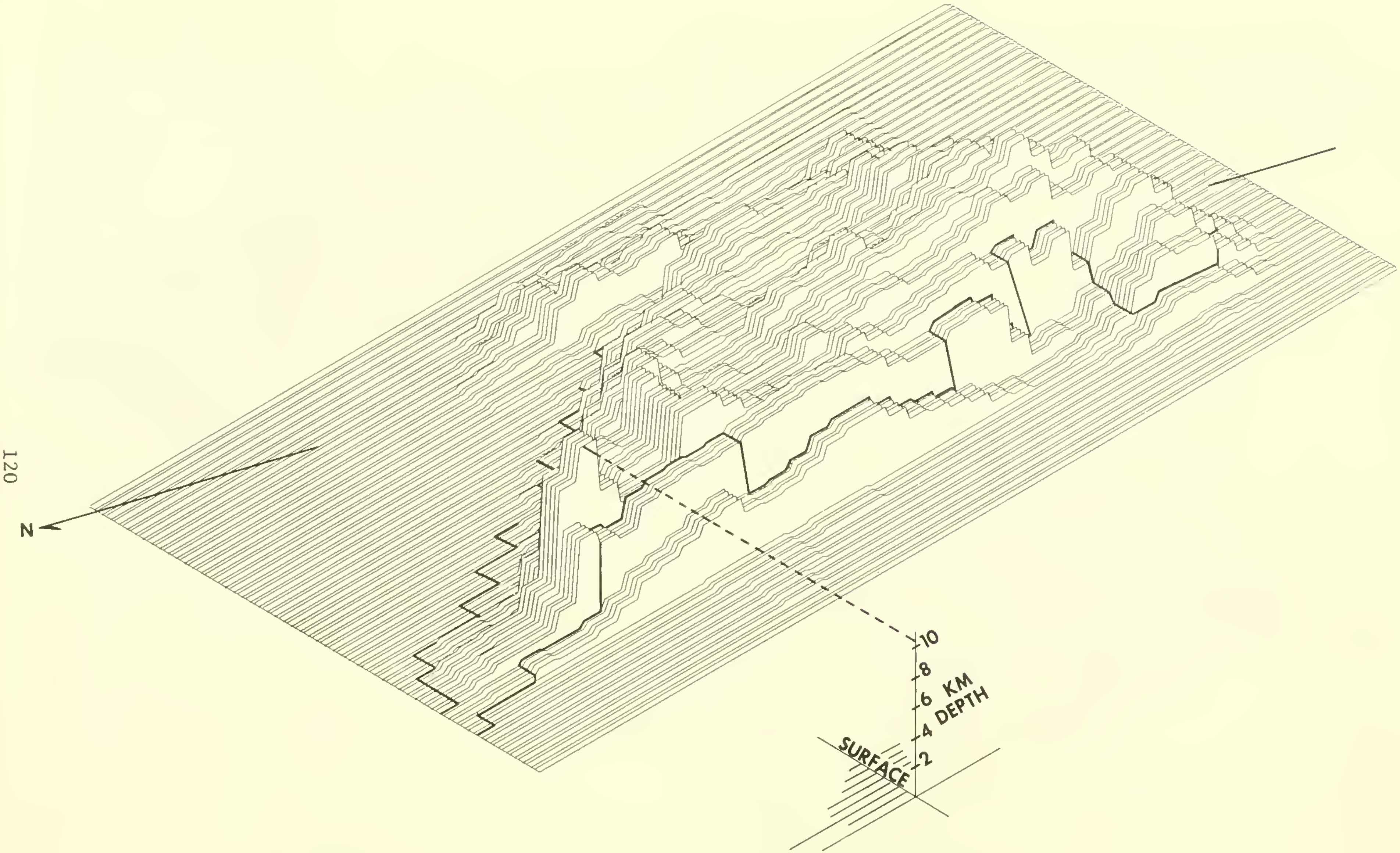

Figure 4b. Same as 4a, view from northeast. Depth scale is east of body.

are either quite thin or have been rendered similar in density to the granites by metasomatism or inclusion of low density dikes and sills associated with emplacement of the two plutons. The anomaly pattern over the Wallamatogus pluton (Fig. 2) suggests inward dipping contacts for the south and southwest parts of the body (Figs. 3 and 4) while the northwest contact is vertical to outward dipping. The Model (Fig. 4) is less than two km thick.

The maximum anomaly gradient zone corresponds remarkably well with the Lucerne pluton contacts except over the northernmost exposures and over the East Blue Hill and Wallamatogus plutons. This suggests that the Lucerne body has predominantly vertical contacts with depth. The northernmost ten km of this granite has no significant associated gravity anomaly indicating that the batholith is quite thin in this region. The deepest zones (greater than 7 km thick), which are somewhat linearly arranged, as well as the main volume of granitic material occurs on the east side of the body (Fig. 4). The significant thicknesses of granite built out under the wallrock on the east side of the model (Fig. 4b) indicate that the Lucerne-wallrock contacts dip very slightly toward the east (about 5°) in the region of maximum east-west surficial extent. The northwest contact is similarly inclined somewhat less steeply (as much as 15° to 20° east of vertical).

The north-south trending elongate anomaly trough over the isoclinal fold nose of the Passagassawakeag gneiss adjacent to the Lucerne pluton along its west side (Fig. 2) indicates the extent of the concentration of largely unexposed pluton called the Stricklen Ridge. Figure 4a shows the body to be generally over three km thick, 1.5 to 3.0 km wide and over 11 km long (about $10^2 km^3$ in volume) widening at its exposed northern end to about 5.0 km in a generally easterly direction. Its east side contacts appear to be near vertical. The west side contacts appear somewhat less steep in general and may dip outward. The influence of low density nonintrusive material within the Passagassawakeag gneiss may be significant, however, so that the apparent outward dip of the Stricklen Ridge body on its west side may not be due to the intrusive.

Two and three dimensional modeling based on gravity maps (Figs. 1 and 2) reveals that the major granite intrusives are relatively thin and equant to tabular in shape with steeply dipping contacts. Steepening of foliation dips in the contact zones of the Lucerne pluton, gravity evidence of upward dragging of modeled bodies adjacent to the Lucerne pluton and the moderate temperature maximum of the relatively narrow Lucerne pluton contact aureole indicate intrusion of the Lucerne magma, at least, as a plastically yielding highly viscous body, probably a solid. The apparent shapes together with structural and contact metamorphic features of the granites and surrounding rocks suggest that all granites, except for the Stricklen Ridge granite, were emplaced at a late stage in the tectonic history of the area.

The Stricklen Ridge pluton intrudes a high grade (sillimanite) gneiss. The gravity evidence (Fig. 2) shows this body to be elongated approximately perpendicular to regional foliation trends, not typical for stress induced accumulations. Also, the fairly steep anomaly gradients associated with this body indicate steep sharply defined contacts, not diffuse ones as might be expected with metamorphic accumulations. Hence, it is felt that the Stricklen Ridge pluton was originally an igneous intrusive which was subsequently metamorphosed and faulted.

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IGNEOUS PETROLOGY OF SOME PLUTONS IN THE NORTHERN PART OF THE PENOBSCOT BAY AREA

Appendix B

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Introduction

The purpose of this appendix is to familiarize the 1974 NEIGC participants with available radiometric age data for granitic rocks to supplement the report by Wones (this Guidebook). The locations with names of granites for which age data are available elsewhere in this Guidebook. As work is currently being carried out on these granites, many of the data are preliminary. The data in Table 1 are so marked to indicate which data are considered highly reliable, etc. Mineral K-Ar and Rb-Sr age data are reported separately from whole rock Rb-Sr data as discrepancies between mineral and whole rock dates are not uncommon.

Constants used for the age calculations are as follows: ^{40}K: λ_e = 0.585×10^{-10}/y, $\lambda_B = 4.72 \times 10^{-10}$/y. ^{87}Rb: $\lambda_B = 1.39 \times 10^{-11}$/y. Standard methods were used for the Rb-Sr chemistry and mass spectrometry (e.g. for previously unreported dates; see also Brookins and others, 1973). The York least squares method has been used for Rb-Sr isochron construction and initial ratio (R_o) determination. Only the preliminary age data are given in Table 1; the complete set of data will be formally published later.

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Table 1: Age Data For Granitic Plutons

Granite	Age (m.y.)	R_o*	Method**	Reference
Stonington	350, 360	---------------	K-Ar B	1
Stonington	341 ± 21	0.7050 ± 0.0006	Rb-Sr WR	2
Oak Point	354	---------------	K-Ar B	1
Oak Point	343 ± 15	---------------	K-Ar B	3
Oak Point	357 ± 1	0.7045 ± 0.0001	Rb-Sr WR	2
Vinalhaven	361 ± 7	0.705 ± 0.001	Rb-Sr Wr	4
Sedgwick	413 ± 15	---------------	K-Ar B	3
Sedgwick	395 ± 15	0.706 ± 0.002	Rb-Sr WR	5
East Blue Hill	387	---------------	K-Ar B	1
East Blue Hill	391	(0.710)	Rb-Sr B	1
East Blue Hill	392 ± 12	0.707 ± 0.002	Rb-Sr WR	5
South Penobscot	360 ± 15	---------------	K-Ar B	3
South Penobscot	393 ± 15	(0.706)	Rb-Sr B	6
South Penobscot	403 ± 15	0.706 ± 0.002	Rb-Sr WR	6
Wallamatogus	379	---------------	K-Ar M	1
Wallamatogus	413	(0.710)	Rb-Sr M	1
Wallamatogus	330	---------------	K-Ar B	1
Wallamatogus	349	(0.710)	Rb-Sr B	1
Wallamatogus	(375 ± 20)	(0.715 ± 0.005)	Rb-Sr WR	6
Mt. Waldo	325	---------------	K-Ar B	7
Mt. Waldo	390 ± 10	0.705 ± 0.002	Rb-Sr WR	6
Lucerne	385	---------------	K-Ar B	1
Lucerne	402 ± 13	0.714 ± 0.002	Rb-Sr WR	6
Cadillac Mtn.	397	---------------	K-Ar B	1
Cadillac Mtn.	398	(0.710)	Rb-Sr B	1
Cadillac Mtn.	369 ± 35	0.7071 ± 0.0015	Rb-Sr WR	8
Somesville	389	---------------	K-Ar B	1
Somesville	419 ± 16	0.7060 ± 0.0026	Rb-Sr WR	8

* R_o = initial $^{87}Sr/^{86}Sr$; () indicate assumed values.

** B = biotite, M = muscovite, WR = whole rock

THE PALEONTOLOGY OF THE PRESENT: LITTORAL ENVIRONMENTS ON A SUBMERGED CRYSTALLINE COAST, GOULDSBORO, MAINE

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Introduction

The frequently quoted edict that "-the present is the key to the past" is a powerful analytical and interpretive tool for the geologist. All too often, however, geologists seem to handle this tool as though it had only one cutting edge or working surface. They collect large volumes of samples, fill many notebooks with data, and organize these data into various, hopefully meaningful, relationships. And then they look for modern counterparts, parallels or analogs. This field trip will illustrate the other approach in handling the tool.

To this end, the central purpose of this trip is to examine the intertidal zone at three localities in the Gouldsboro Peninsula: Sand Cove, Corea, and Little Moose Island (Schoodic Point). The substrate, and conditions in the water column, are different at each of these localities. Differences in the biota at the three localities are expressions of the differences in the opportunities presented, which themselves are ultimately dependent on the physical conditions.

A casual glance at the road log will reveal what appears to be a needless amount of back-tracking, from Stop 4 to Stop 5. This is made necessary in part by the one-way road around Schoodic Peninsula, and chiefly by constraints of the tide, for we cross to Little Moose Island only when the tide has ebbed sufficiently.

Stops 1, 2, and 5 are in the Schoodic Peninsula portion of Acadia National Park; Stops 3 and 4 are on private property. Please help us conduct a field trip in which due respect is given to all property, public and private alike.

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Itinerary

Mileage

0	Assemble in Frazer Point picnic area. This is reached by taking Rt. 186 east from Winter Harbor and then turning right at the Gateway Motel (0.6 miles from Winter Harbor). The entrance to Frazer Point is 1.6 miles south of the Gateway Motel.
0.2	Leave Frazer Point picnic area and turn right (start one-way road).
0.4	First open view to right, across to Mt. Cadillac.
0.9	Good view to right, shows cliffs on Ironbound.
1.3	Best view, opposite Mark Is., showing Mt. Cadillac, the Cranberry Islands, Ironbound, Turtle, Porcupine with glacial plucking from lee side, etc.
2.5	Entrance on left to Schoodic Head road (end mileage log). Proceed up this road to parking area, and continue on foot to Schoodic Head.

Stop 1. Schoodic Head.

The purpose of this stop is to examine the regional setting. A brief walk over the summit of Schoodic Head (elevation:440') brings one to a witness post and a panoramic view, north,

east, and west, of the Gouldsboro Peninsula, the southwestern tip of which is called the Schoodic Peninsula.

The bedrock of the Gouldsboro Peninsula consists of a number of different kinds of granite, intruded by basalt and felsite dikes. Fractures and joints in the bedrock have largely influenced the results of glacial and marine erosion.

The hills to the north, approximately 16 miles distant, are monadnocks of granite, part of the Tunk Lake pluton (Chapman, 1968, p. 386), and having a composition and texture generally similar to the medium- and coarse-grained granites that form most of Mount Desert Island (Chapman, 1968, p. 388; 1970, p. 41). Schoodic Mountain, on the left (elevation: 1069'), has a pronounced asymmetry due to glacial plucking from the south-facing, lee side. The erosion surface above which these and other resistant monadnocks rise is the New England Peneplain, and is probably of subaerial origin (Raisz, 1929, pp. 138-139).

To the east the peninsula is bounded by Gouldsboro Bay. The nearer small points of land with intervening harbors, which make up the irregular submerged coastline of the peninsula include Spruce Point, Birch Harbor, Prospect Harbor, Cranberry Point, and Corea. The points of land beyond Gouldsboro Bay include Dyer Neck, Petit Manan Point, and Grand Wass Island. The lighthouse flashing on an island about eight miles distant is Petit Manan Light.

Immediately to the west and northwest is Winter Harbor. A granite promontory called Grindstone Neck, and Turtle Island serve to separate Winter Harbor and its approach from Frenchman Bay, which forms the main western boundary of the Gouldsboro Peninsula. Beyond Frenchman Bay, the western horizon is dominated by Mount Desert Island, with its east-west range of granite monadnocks, the highest of which, Mount Cadillac, rises to an elevation of 1530', and the nearest of which is Champlain Mountain.

Among the islands in Frenchman Bay to the northwest are Ironbound and Bald Porcupine, both of which show escarpments facing generally south. These cliffs are formed by a combination of rock structure and glacial action. The islands in the northern half of Frenchman Bay, including Ironbound and Bald Porcupine, are largely capped by a resistant diorite sill, intruding clastic Devonian sediments of the Bar Harbor group (Chapman, 1970, p. 41). On cooling, the sill developed columnar jointing, and where later deformation caused the sill to dip gently to the north, as at Ironbound and Bald Porcupine, weathering and erosion controlled by the nearly vertical joints resulted in the steep south-facing palisades. Glacial

plucking from the south, lee faces of these cliffs served to accentuate them. Because of their location well within Frenchman Bay, it is doubtful that wave action has done more than simply maintain these cliffs by undermining the weaker Bar Harbor sediments beneath the diorite columns (Chapman, 1970. pp. 36-37).

Schoodic Head is composed of a fine-grained granite, similar to that found on the southern part of Mount Desert Island (Chapman, 1970, p. 91). This fine-grained granite will be discussed further at Stop 2.

The north-facing stoss slopes of resistant bedrock hills, such as Schoodic Head, are typically smoothed and polished by glacial abrasion. Two examples of preserved glacial polish, one a few feet south of the witness post, the other near the summit benchmark, will be seen.

The south-facing lee slope of Schoodic Head is a steep escarpment with a talus of large granite blocks at its base, and was formed by glacial plucking and quarrying of the jointed bedrock.

This typical glacial erosional sculpting of stoss and lee slopes finds expression on many of the hills and mountains of the Couldsboro Peninsula and environs. As mentioned above, it is seen on the monadnocks to the north, and is well developed on the hills and mountains of Mount Desert Island to the west.

3.0 Resume road log on one-way road, and continue to point opposite marsh on left between Big Moose and mainland. The road here is located on one of two tombolos that connect Big Moose Island to the mainland.

3.4 Intersection road right to Schoodic Point, two-way traffic.

4.0 Parking area.

Stop 2. Schoodic Point

(Note: Although the southern tip of Big Moose Island, which is now connected to the mainland by two tombolos, is usually called Schoodic Point, as indicated by road signs, the correct geographic location of Schoodic Point is the southern end of Little Moose Island, to the east (Chapman, 1970, p. 87). However, we will bow to custom and sustain this nomenclatural error.)

This stop affords a closer look at the rocks that form most of Gouldsboro Peninsula. Chief amongst them are the medium - to fine-grained granites of the granitic-granophyric phase of the Bays-of-Maine igneous complex, a generally stratiform and lopolithic intrusive body, gabbroic below and granitic above, that invaded Silurian and early Devonian volcanics and sediments, and was itself later intruded by circular stocks of

coarser-grained granite, one of which will be seen at Stop 4 (Chapman, 1962, pp. 883-886; 1968, pp. 386-387). The particular granite exposed at Schoodic Point is of the fine-grained variety, and may extend uninterruptedly beneath the mouth of Frenchman Bay westward to Sutton Island and Southwest Harbor on the southern half of Mount Desert Island (Chapman, 1970, p. 91).

Felsite porphyry dikes, exhibiting flow structures, intrude the granite, and are perhaps contemporary with the younger plutons of coarser-grained granite (Chapman, 1970, p. 26). One such felsite dike will be seen at the western edge of Schoodic Point, and another will be seen later at Stop 5.

The granites here and elsewhere are also intruded by numerous basic dikes, of which four distinct groups have been discerned (Chapman and Wingard, 1958, pp. 1193-1195; Chapman, 1970, p. 26). At least two of these four groups are evident on Schoodic Point. An example of a multiple dike will be seen near the eastern edge of the point.

In examining these rocks at Schoodic Point, one could do no better than to follow the splendid self-guided tour given by Dr. Chapman (1970, pp. 91-100), and this we shall, with slight modifications, do. His description is such a model of clarity that even a pair of surf-dazed paleontologists were able to follow it.

Indicators of the direction of flow of the glacier are also evident at Schoodic Point, and confirm the general regional north-to-south movement of the ice (Raisz, 1929, pp. 157-159). These indicators include striae, crescentic gouges, and crescentic fractures.

4.2 Natural sea wall on right, for 0.3 mile.

4.5 Intersection, back on one-way road.

7.2 Park boundary and beginning of two-way traffic, at Wonsqueak.

9.2 Intersection with Rt. 186, at Birch Harbor.
Turn right on Rt. 186.

11.3 Intersection, Rt. 195 joins Rt. 186 from left, at Prospect Harbor.

11.5 Intersection, turn right on Rt. 195, at Prospect Harbor.

12.5 Sand beach on right. Park cars.

Stop 3. Beach in Sand Cove

Part of the rock debris resulting from the erosion of the rocky headlands is carried out into deeper water; part is

carried into the bays and coves where it may be deposited forming beaches. These beaches are constantly being modified by the deposition of new material and the removal of dry sand, particularly from the upper part of the beach, by the wind. Sand is thus carried inland above the strand-line forming dunes and helping to infill salt marshes. During storms, material from such beaches may be moved into deeper water.

There is such a beach at the head of Sand Cove, a rock-bounded cove approximately 1500 yards east of Prospect Harbor. This beach shelves gently and is made largely of quartz sand (the results of a more detailed study of sand samples from this beach will be distributed at the meeting).

The beach is approximately 400' long in roughly an east-west direction, and its width varies with the stage of the tide. A narrow backshore rises to a dune ridge, behind which a tidal swale occurs, draining into the sea through a boulder field, then over the beach sand, at the eastern edge of the beach. The beach is bounded, east and west, by granite outcrops. The granites are distinctly different on each side. On the west the granite is medium-grained, pink, with two feldspars. On the east the granite exhibits a wide range of texture, is red, and is dominated by one feldspar. The map (C & GS Chart 305) suggests a structural cause for the contrast in granites: a narrow linear arm of Gouldsboro Bay, called Grand Marsh Bay, runs nearly into Sand Cove, and may indicate a fault zone.

The sessile benthos to be seen on Little Moose Island and on the rocks bounding Sand Cove is absent from the sand beach because it cannot establish itself on an unstable substrate. Instead, the unconsolidated, loose nature of the substrate creates opportunities for infauna. The infauna here is not highly diversified and is exemplified by the Atlantic Jackknife clam, the soft-shell clam, and Moon snails, represented by their shells on the beach; amphipods (probably beach-fleas) represented by their burrows high up on the beach; and worms of some kind represented by accumulations of ejected sediment resembling coiled rope in their form.

Some of the organic remains on the beach have been thrown up from deeper water by the waves. Among these are slipper shells, sand dollars, and crabs.

There are also biogenic structures on the surface of the sand: Footprints of gulls and raccoons, and locomotion trails made by snails. All these are potential trace fossils.

Sedimentary structures, braided stream channels where a stream crosses the eastern end of the sand beach, and rill-marks, could be preserved as primary structures in rock.

14.7 Village of Corea. Continue straight ahead, following signs for "Harbor Grocery".

15.2 Harbor Grocery, Corea. Be careful in parking so as to leave as much room as possible for other cars of grocery customers to park and turn. Lunch.

Stop 4. Corea

The harbor at Corea is bounded on the west and north by a dominantly granitic, low-lying shore-line, and on the east by an island, also granitic, connected to the mainland by a tombolo. This tombolo consists of granite pebbles and cobbles on the east side where it faces Gouldsboro Bay, and a salt marsh on the west where it faces Corea.

The bedrock belongs to the Corea pluton, a coarse-grained, two feldspar, porphyritic to sub-porphyritic granite with rapakivi texture (Chapman, 1968, p. 388). It is closely similar to the Tunk Lake pluton, which includes Schoodic Mountain, to the north.

There is a small sand beach just south of the Harbor Grocery on the west side of Corea Harbor. The beach material shows gradation from angular, weathered granite pebbles, to coarse arkosic sand. The grains in the arkosic sand are still generally angular. This material is interpreted as having been derived from the local bedrock with very little transport. Hence it may be considered as a granite wash (the results of a more detailed study of sand samples from this beach will be distributed at the meeting).

The sand here shows an upper zone, approximately 3 or 4 inches thick and reddish in color, that is oxidized. In contrast, the underlying sand is darker and is presumed to be somewhat richer in organic matter (Bradley (1957, p. 652) described a similar situation in the sandy sediment in Sagadahoc Bay, Georgetown, Maine). Abundant castings of ejected sediment, similar to those seen at Sand Cove, display this darker sediment. It is probable that much of the organic matter is fecal, contributed by the 'worms' responsible for the castings seen on the surface. Organic activity is therefore presumed to be a factor in the reworking of the sediments here.

Other organic remains include shells of the soft-shell clam, the foraminiferan Elphidium, ostracodes, as well as a number of trails.

18.9 Intersection, Rt. 195 and Rt. 186, in Prospect Harbor. Turn left and continue on Rt. 186.

21.3 Birch Harbor. Continue on Rt. 186 toward Winter Harbor.

23.1 Turn left at Gateway Motel toward Schoodic Section, Acadia National Park.

26.3 Intersection. Bear left toward Birch Harbor and Winter Harbor.

26.9 Entrance, on right, to Blueberry Hill parking area. Park cars, walk back on road approximately 0.3 mile. and cross tombolo to Little Moose Island.

Stop 5. The south end of Little Moose Island

Be careful of your footing in crossing over to the island. Wet boulders covered by wet bladder wrack tend to be exceedingly slippery.

Approaching the main body of the island, notice the erosional exposure, on the right, of an 8' thick layer of glacial till deposited on the north side of Little Moose. Except for thin and isolated patches of peaty soil, the south side of the island consists of exposed bedrock.

Further caution is urged in approaching the south end of Little Moose Island, due to a field of large tumbled boulders, and the wet rock surfaces near the water's edge.

The bedrock of the south end of Little Moose Island chiefly consists of a medium-grained granite belonging to the granitic-granophyric phase of the Bays-of-Maine igneous complex (Chapman, 1962, pp. 883-886).

A wide felsite porphyry dike, similar to one seen at Stop 2, intrudes the granite. Both the granite and the felsite dike are themselves later intruded by basic dikes, one of which is a multiple dike with internal chill borders.

Low cliffs and "holes" have been eroded along the basic dikes and fracture zones in the granite. The rock surfaces exposed here support a diversified biota. Its most conspicuous component is sessile benthos which has been able to establish itself here because the substrate is solid.

The biota shows a vertical zonation in, and immediately adjacent to, the intertidal zone. Six zones are recognized (Carson, 1955 p. 30-33, 39-123; Edey, 1972, p. 61; Kingsbury, 1970, p. 25-53):

The Black Zone
The Periwinkle Zone
The Barnacle Zone
The Rockweed Zone
The Irish Moss Zone
The Laminarian Zone

The distribution of organisms within the intertidal zone is determined by a number of environmental factors amongst which the relative exposure to sea-water and air, and the related desiccation, temperature fluctuations, and exposure to fresh-water during rain, as well as competition with other organisms are perhaps the most important. Those organisms that need the least

immersion in sea-water and hence are more tolerant to exposure to air occur in the highest zones; tolerance to exposure to air thus imposes an upper limit to the occurrence of sessile benthos. Lower limits, on the other hand, are imposed by competition from organisms that are better adapted to the conditions in the lower zones.

The Black Zone is so called because of a film of blue-green algae on the rocks above the limits of the highest Spring Tides. The algae in this zone are moistened by salt spray.

The Periwinkle Zone lies below the limit of the highest tides and above the limit of average high tides. It is thus covered by seawater twice in each lunar month, during the Spring Tides. It is populated by rough periwinkles, air-breathing viviparous gastropods which graze on the algae growing on the rocks in this zone and, during Spring Tides, on the algae in the overlying Black Zone. Because of their near adaptation to terrestrial life, rough periwinkles climb the rocks to higher levels to escape drowning during the highest, Spring tides.

The Barnacle Zone is a conspicuous white zone caused by the occurrence, and local extreme crowding, of white balanoid barnacle shells attached to the rocks. This zone receives the most intense battering by the waves and barnacles are the only sessile organisms that have been able to establish themselves here. Barnacles are tolerant of considerable exposure to air but need approximately one hour submergence during a tidal cycle. While submerged, they feed on small organisms and particles of suspended organic matter, which they collect on their legs. Barnacles are fed on by dog whelks, worms, and fish.

Although balanoid barnacles can survive a greater degree of immersion, they are not found in any numbers below the barnacle zone. A number of factors may contribute to their virtual exclusion from lower zones: First, a greater degree of submergence may lead to greater contact with, and hence vulnerability to, predators such as dog whelks. Second, adult barnacles, already established on the rocks high up in the intertidal zone, may attract barnacle larvae by means of a chemical 'scent' released into the water. Third, competition from creatures adapted to the lower zones may exclude barnacles from those zones.

The Rockweed Zone is marked by the dense growth of the brown algae called Rockweed or Bladder Wrack. The upper limit of this zone is determined by the Rockweed's need for at least an hour of submergence during each tidal cycle; the lower limit of the zone is determined by competition from Irish Moss that is better adapted to the somewhat dimmer illumination in deeper water. The Rockweed Zone is exposed to air, on the average, for half the time.

The Rockweed provides hiding-places for a number of animals; mussels, common periwinkles, smooth periwinkles, limpets, crabs, and worms.

The Irish Moss Zone is distinguished by the red alga called Irish moss and is exposed only during the lowest, Spring Tides. It is populated by sea-urchins, star-fish, and crabs, as well as the green alga Ulva or sea-lettuce.

The Laminarian Zone lies below the lower limits of the lowest Spring Tides and hence is never exposed to the air. Laminaria is a brown alga, represented here by the Oarweed and kelp. This zone is populated by sponges, sea-anemones, jellyfish, worms, sea-urchins, starfish, crabs, and tunicates.

In addition to the six zones noted here, there is a number of tidal pools at this locality. These pools are situated in basins and depressions developed in the fractured zones in the dominantly granitic bedrock in the intertidal zone. Salt water is trapped within these basins and depressions when the tide ebbs. At flood tide, circulation is restored, nutrients are brought in, and wastes flushed out.

Because the biota in the tidal pools is never exposed to the air, it bears a close relationship to the biota of the Irish Moss and Laminarian Zones. On the other hand, the ponding of water in the pools during ebb tide causes the environment of the tidal pools to differ from that of the open water. The water in a tidal pool is more rapidly warmed by the sun, particularly during the summer. There is, therefore, a more strongly marked diurnal temperature cycle in the pools. Metabolic processes of the enclosed biota may lead to changes in the composition of the ponded water. Furthermore, the addition of rainwater may markedly reduce the salinity of the pool. The flushing action of a flood tide may thus cause extremely rapid changes in the temperature and composition of the water enclosing the biota of a tidal pool.

The tidal pools at Little Moose Island contain a variety of algae: red algae include Pink Rock-Crust, a pink encrusting alga; and Branching Corallina, an erect ramose lime-secreting alga. Green algae include Cladophora which grows in silky tufts; and Ulva, the sea-lettuce, which forms bright green membranous sheets. Brown algae are represented by the Oarweed and kelp. A number of animal phyla are represented by encrusting sponges, sea-anemones, chitons, whelks, periwinkles, limpets, mussels, sea-urchins, star-fish, and tunicates.

LATE WISCONSIN AND HOLOCENE GEOLOGICAL BOTANICAL, AND ARCHAEOLOGICAL HISTORY OF THE ORONO MAINE REGION

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Introduction

This trip is designed to observe various aspects of the Late Wisconsin and Holocene history of the Orono region. Research in this region is ongoing and involves both undergraduate and graduate students as well as staff of the departments of Anthropology, Botany and Plant Pathology,and Geological Sciences,and the Institute for Quaternary Studies.

Glacial Geology

The glacial deposits of the region are characterized by clay-rich ground moraine and extensive prominent esker systems left directly by the Late Wisconsin Laurentide Ice Sheet that moved over the area to the southeast (Prescott, 1966).

These deposits are stratigraphically overlain by glacio-marine clay silt and sand of the Presumpscot Formation (Bloom, 1960). This formation contains a cold-water flora and fauna characteristic of the modern sub-arctic (Bloom, 1960).

The margin of the Late Wisconsin Laurentide Ice Sheet, retreating from a terminal position on the continental shelf, approximately coincided with the present Maine coast by 13,500 years B.P. (Borns, 1973). The recession of the margin continued inland accompanied by a marine submergence which flooded the lowlands of central Maine and formed estuaries in the Penobscot Valley as far inland as East Millinocket and in the Kennebec Valley as far as Bingham (Goldthwait, 1949; Borns, 1963).

The submergence at the present position of the coast began approximately 13,500 years B.P. Subsequently the sea flooded glacially depressed central Maine and as crustal rebound exceeded eustatic sea level rise the sea drained to the position of the present coast by approximately 12,100 years B.P. Thereafter relative sea level continued to fall to approximately - 180 feet by 10,000 years B.P. (Milliman and Emery, 1968). Subsequently the sea level has risen to its present position.

Botany

Three radiocarbon-dated pollen diagrams have been prepared from Maine. These are from Mud Pond, near Waterville (Davis, 1969); Moulton Pond (Bradstreet, 1973) and Holland Pond (Davis, in prep.) in the Bangor-Orono area. Only Moulton Pond includes the complete late glacial record.

Starting at approximately 13,500 years B.P. tundra vegetation occupied the area. The landscape was particularly barren until about 12,000 years B.P. when the lowlands emerged from the early seas of sub-arctic character.

Transition to forest started about 10,500 years B.P., ending with the establishment of the white pine and birch-dominated forests by 9,700 years B.P.; oak also became important. This association was joined by hemlock and northern hardwoods about 7,000 years B.P.; pine declined to low values at this time at Moulton Pond, but remained common at Holland Pond until 4,000 years B.P. Temperate taxa peaked about 5-3,000 years B.P. The most recent 1 or 2 thousand years is characterized by increases in spruce.

Archaeology

The earliest known occupation of the area was at the Hirundo site on Pushaw Stream, a tributary of the Penobscot and probably dates to about 7,000 B.P. The site is found in flood silts resting directly above an eroded till surface. The first major occupation, however, is the Laurentian Tradition dated at this site to at least 4,300 and possibly earlier. In addition to the Laurentian are found later occupations coming up to the historic period. Although vertical separation of the cultural components is not as clear as desirable, there is some separation on a horizontal basis. The Hirundo site is located on the only stretch of rapid water on Pushaw Stream. As such, there were annual runs of alewife, salmon, and possibly shad, which could be taken in the shallow waters with relative ease. Because of the acidic soils preservation of organic remains is practically non-existent.

To date, approximately three percent of the site has been excavated. It has revealed some most useful data and has provided for the first time in the interior of the state some provenience information on previously "floating" collections.

The Hirundo Archaeological Project is designed to document and explain prehistoric' man's adaptation of Maine's interior. The project is conceived as a multi-disciplinary one involving Quaternary scientists of the Institute for Quaternary Studies. Grants from the National Geographic Society have sponsored field and laboratory work.

While interest in the prehistory of Maine is not new the nature of much of the earlier work is such that little useful data are available, especially in the important areas of man's adaptation to past environments

in select ecosystems. The analysis of the Holland Pond core, taken about six miles from the Hirundo archaeological site, is providing insights into past vegetation communities, while goelogical and limnological studies are providing data on bog formation, and past drainage patterns, all of which are critical to a proper understanding of prehistoric man.

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Itinerary

Mileage	
0	Assembly point for trip is in the parking area west of Field House on U.M.O. campus. 8:30 A.M. Drive west on campus to Rt. 2A.
0.2	Turn left on College Ave. (Rt. 2A).
1.6	Turn right in Orono center on Forest Ave.

6.2	Cross Essex Street. Continue west
7.7	Turn left on Pushaw Road.
9.0	Stop 1. Turn left into Mahar's gravel pit. This borrow pit exposed a complete cross section of an esker, the till beneath and the fossiliferous marine sediments on top. Return to Pushaw Road and turn right.
10.3	Turn right on Forest Ave.
11.8	Cross Essex Street. Continue east.
16.0	Turn left on Stillwater Ave.
17.3	Turn left on Bennoch Road.
17.8	Turn left on Kirkland Road.
22.6	Stop 2. Park along the road. Caribou Bog was developed primarily upon glacio-marine silts and clays after the area emerged about 12,100 years B.P. Here the postglacial vegetational sequence will be discussed. Turn around and drive east on Kirkland Road.
22.9	Turn left on West Old Town Road.
27.9	In West Old Town turn left.
28.1	Stop 3. Turn right into Hirundo Nature Preserve and park where directed. The Hirundo Archaeological Site will be discussed in reference to its people, location and environment at time of occupation approximately 7000 years B.P. Return to main road and turn right.
28.3	In West Old Town turn right.
33.3	Bear left and continue towards Bennoch Road.
35.1	Turn right on Bennoch Road.
35.6	Turn left on Stillwater Ave.
36.1	Turn right on College Ave. (Rt. 2A)
37.1	Turn left on to U.M.O. campus.

BEDROCK GEOLOGY OF MOUNT DESERT ISLAND

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Introduction

Mount Desert Island, the largest bedrock island on our Atlantic coast, lies on a long belt of lower and middle Paleozoic rocks that trends roughly parallel to the Maine coastline. The rocks of this belt were derived principally from magmatic sources over a geologically long period of time (probably Ordovician to Devonian). The older representatives, originally stratified pyroclastic materials and interlayered lava flows, constitute the (1) Ellsworth Schist, (2) Cranberry Island Series, and (3) Bar Harbor Series. The younger representatives, principally gabbroic and granitic intrusive rocks, constitute the (1) Bays-of-Maine igneous complex and (2) Maine coastal plutons.

In its central portion, Mount Desert Island is underlain by granite (probably Devonian) which in turn is surrounded by a nearly continuous fringe of older rocks (fig. 1). The granite core is composed of two distinct units, each representing one of the Maine coastal plutons. The larger and older body, the Cadillac Mountain pluton, is composed of coarse-grained hornblende granite. The smaller, younger body, the Somesville pluton, is composed largely of medium-grained biotite granite.

This dual pluton is bordered on the west and northwest by a gabbroic phase of the Bays-of-Maine igneous complex (probably Silurian-Devonian) out beyond which lies schist, gneiss, and quartzite of the Ellsworth Schist (probably Ordovician). The fringe rocks along the northeastern, eastern, and southern contacts consist largely of contact metamorphosed, well-bedded siltstone and sandstone of the Bar Harbor series (probably Silurian). The southern part of the island is made up of weakly metamorphosed volcanic tuff, lava flows, and intrusive felsite of the Cranberry Island Series (probably Silurian) and a granitic phase of the Bays-of-Maine igneous complex. A number of minor satellitic bodies of granite with arcuate form (fig. 1) represent ring-dikes.

Rock Units

Ellsworth Schist--The term Ellsworth Schist is now being used to include certain rocks on Mount Desert Island and nearby Bartlett Island that Shaler (1889) called the Bartlett's Island series. Though the latter term is the older and should have perhaps maintained priority, it has now fallen into disuse. The age of the Ellsworth Schist is probably Ordovician. It has been considered pre-Middlle Silurian (Smith et al, 1907; Chapman and Wingard, 1958).

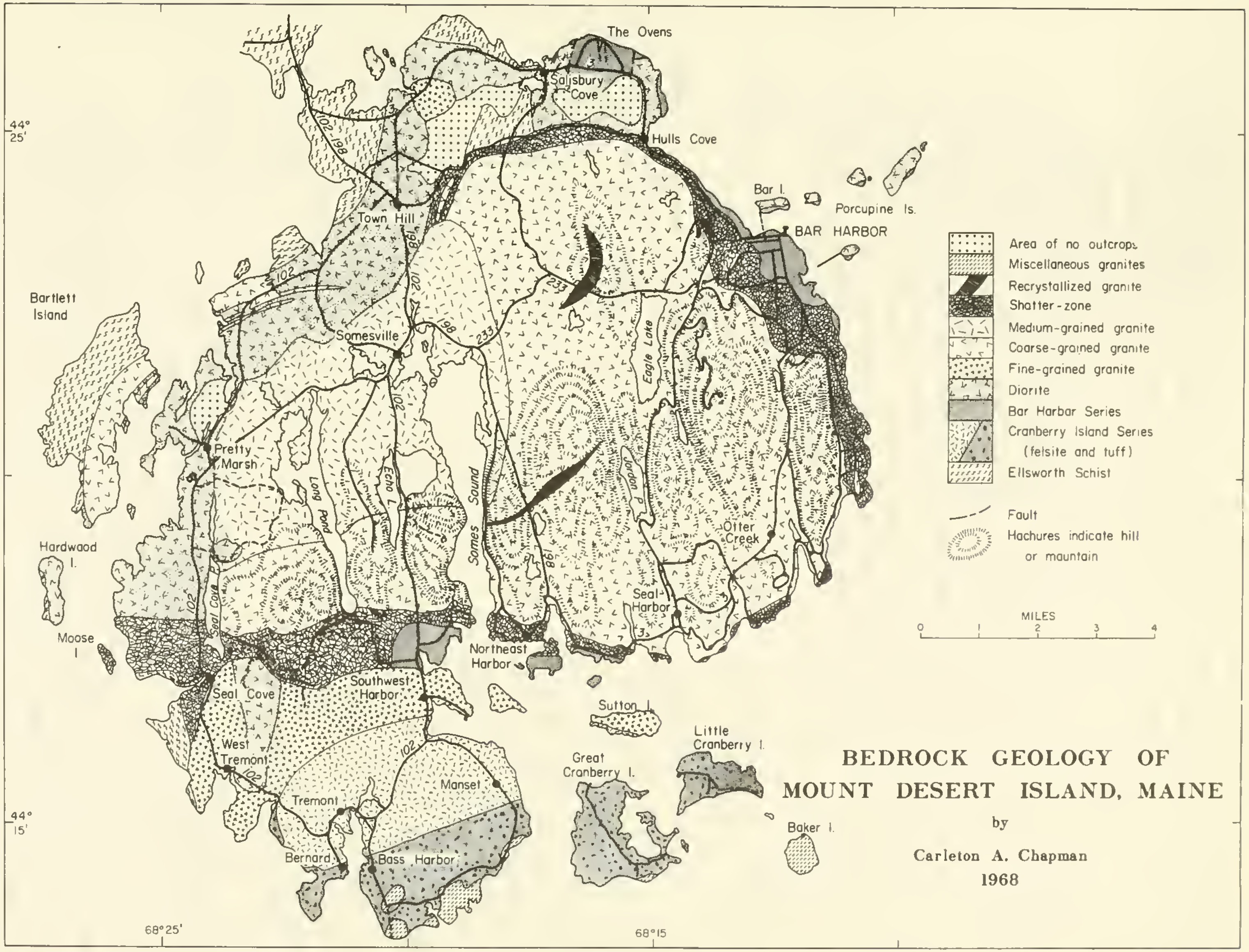

Figure 1. Bedrock geology of Mount Desert Island, Maine.

Reproduced from Chapman (1970) with permission of the Chatham Press, Inc.

The rocks of the Ellsworth Schist are generally quartzofeldspathic with varying amounts of chlorite, muscovite, epidote, biotite, and actinolite and are represented mainly by greenish and grayish chlorite and mica schist, interlayered with smaller amounts of (1) light to dark green amphibole schist and amphibolite, (2) quartzite and feldspathic quartzite, and (3) fine-grained quartz-feldspar gneiss.

Originally the material probably represented water-laid tuffs and some lava flows with compositions largely in the andesite to quartz latite range. Smaller amounts of pelitic and feldspar-poor rocks are found. As a result of low-grade to middle-grade metamorphism the rocks have been well recrystallized with the loss of most original textures and structures. A true foliate structure is generally present, and quartz segregations (lenses, layers, veins) may be abundant and widespread. Small folds and crinkles are common and generally trend N-S to NE-SW. Lineation has many variants and is nearly universal. Faulting does not appear to be common or extensive.

Cranberry Island Series--The Cranberry Island Series (Shaler, 1889) is considered of Middle or Late Silurian age (Chapman and Wingard, 1958). The stratigraphic relation of this series to the Ellsworth Schist cannot be determined directly from field study. This series, however, appears to correlate with the Castine Formation (Chapman and Wingard, 1958) which has been shown to rest unconformably on the Ellsworth Schist near Castine, Maine (Wingard, 1958).

The series originally was composed principally of well-bedded tuff interlayered with lava flows and locally injected by felsite. Compositionally the material is largely andesitic to quartz latitic with smaller amounts of rhyolitic and basaltic rock.

The series appears moderately deformed and layers dip at low to high angles. Locally a conspicuous foliation has developed and may cut bedding at small to moderate angles. The metamorphism was generally weak, and original volcanic textures and structures are still well-preserved. Lineation is locally conspicuous; but faulting, though difficult to detect, does not appear to be common.

Bar Harbor Series--The Bar Harbor Series (Shaler, 1889) may be of Late Silurian age but this matter needs further study. It is represented predominantly by well-bedded brown, gray or green metasiltstone and quartzite. Bedding is thin, commonly graded, and very regular. Cross-bedding and minute cut-and-fill structures are not uncommon. The rocks are highly feldspathic and the grains relatively angular. The beds are believed to represent, in part, water-laid volcanic materials (Chadwick, 1944; Chapman, 1962b and 1969; and Metzger and Bickford, 1972).

At the base of the series is a conglomeratic member (Chapman, 1957) which rests unconformably on weakly foliated volcanic rocks (presumably Cranberry Island Series) and carries pebbles and cobbles of the underlying material.

Over large areas these rocks show horizontal to gently dipping beds. The sharpest flexures are local and apparently associated with movements along steep faults. Normal faults of small extent and displacement are particularly common, and many appear directly related to cauldron subsidence.

Bays-of-Maine igneous complex--This complex (Chapman, 1962a), composed predominantly of layered gabbroic rocks and granites, extends at least from the west side of Pendbscot Bay to well into southern New Brunswick, a distance of 175 miles. The total length may exceed 300 miles; and although the complex may be somewhat discontinuous, it trends roughly parallel to the regional structure.

For simplicity the complex is subdivided into an older, gabbroic phase and a younger, granitic-granophyric phase. It is remarkably similar petrographically and structurally to many other bimodal stratiform complexes (Bushveld, Duluth, etc.). It was emplaced largely within and beneath a thick pile of volcanic and associated stratified rocks (Ellsworth Schist, Cranberry Island Series, and Bar Harbor Series) in late Silurian to middle Devonian time (Chapman, 1962a).

The early or gabbroic phase is highly variable, consisting of norite, gabbro, quartz gabbro, ferrogabbro, diorite, and quartz diorite. Layering (rhythmic, graded, and cryptic) and igneous lamination are common; and on Mount Desert Island the layered sequence may exceed 4000 feet in thickness. On Mount Desert Island perhaps the most common rock type, as well as the average compositional type, representing this phase falls close to the diorite-gabbro boundary (shown as diorite in figure 1).

The younger, granitic-granophyric phase forms an extensive sheet or capping on top of the gabbroic phase and beneath a roof of volcanic material (Cranberry Island Series). The rocks of this phase are medium to fine grained (mostly fine grained on Mount Desert Island) and range compositionally from granite to quartz diorite (fine-grained granite of figure 1). The mafic content (biotite and/or hornblende) generally does not exceed 5 or 10 percent of the rock. Biotite is more characteristic of the granitic textured variety whereas the granophyric variety is generally hornblende bearing. The granites of this phase of the complex should not be confused with those of still younger plutons, next to be described.

Maine coastal plutons--Cutting both phases of the Bays-of-Maine igneous complex and the older stratified rocks are numerous, large, round, stock-like bodies of granite referred to (Chapman, 1968) as the Maine coastal plutons. Some of these have been dated as Early to Late Devonian (Faul et al, 1963). The plutons are 5 to 10 miles across and are composed mostly of medium-to coarse-grained biotite granite. Hornblende granite is much less common.

Two of these plutons occur on Mount Desert Island. The older, Cadillac Mountain pluton is composed of a one-feldspar, hornblende granite with a coarse-grained hypidiomorphic granular texture (coarse-grained granite of figure 1). Much of the rock shows a typical cumu-

late texture. Alkali feldspar (perthite) occurs as closely packed subhedral to euhedral grains with quartz and small amounts of mafic minerals filling the interstices. Commonly more or less quartz is intergrown with the feldspar. The country rock marginal to this pluton has been severely shattered for as much as half a mile from the contact, thus forming a nearly continuous zone of breccia (shatter-zone of figure 1).

The younger or Somesville pluton, composed of biotite granite, cuts out a large portion of the older Cadillac Mountain pluton. The rock is largely a medium-grained, allotriomorphic granular, biotite granite (medium-grained granite of figure 1) with two feldspars (perthitic microcline and sodic plagioclase). The grain is coarsest in the southern and eastern parts, and a fine-grained phase constitutes a crescent-shaped mass in the central area. Over much of the pluton the rock is subporphyritic, and this texture is most prevalent in the finer grained varieties.

Emplacement of the plutons

The mode of emplacement of the two plutons on Mount Desert Island is considered (Chapman, 1953) to be by ring-fracturing and cauldron subsidence. The same explanation appears to hold for the other Maine coastal plutons (Chapman, 1968). The supporting evidence is as follows: (1) sub-circular outline of the plutons; (2) smoothly-curved, sharp boundaries; (3) paucity of granitic apophyses; (4) ring-dikes passing into shatter-zones and breccia; (6) contact zone of brecciation (shatter-zone of figure 1) around the plutons; (7) web vein patterns and web breccia patterns (Chapman, 1968) in adjacent country rock; (8) external centripetal dip pattern (Chapman, 1953) in adjacent country rock; and (9) chilled contact zones implying rapid ascent of magma.

In the case of the Cadillac Mountain pluton, the subsiding block appears to have been completely detached from the country rock and to have dropped entirely below our present level of observation. After the magma crystallized, further subsidence in the northwestern part of the pluton disturbed the granite and localized its recrystallization along arcuate belts (recrystallized granite of figure 1). Nearly contemporaneously subsidence created a number of ring-dikes in the western part of the island and vicinity. Subsequently the Somesville pluton formed, but in this case the subsiding block was not completely isolated from the rest of the country rock by arcuate fractures. It remained attached along the northern and western sides and subsided mostly along the southern and eastern sides. This differential movement caused the block to tilt to the southeast as it flexed in hinge-like fashion along the attached sides. Magma entered the gap to the southeast and spread up over the depressed portion of the block to form an incomplete ring-dike of medium-grained biotite granite. A second subsidence and flexing soon followed, and a new and smaller gap in the central and perhaps incompletely crystallized portion of the ring-dike opened. This led to the formation of a second ring-dike of crescent shape and fine-grained texture within the first. This younger intrusion is not shown on the geologic map (fig. 1).

Pattern of consolidation in the plutons

On a textural-structural basis the Cadillac Mountain pluton may be divided into three concentric and more or less distinct zones (Chapman, 1969): (1) contact zone, (2) marginal zone, and (3) central zone. The outermost or contact zone (several hundred yards wide) is composed of the finest-grained hornblende granite. At its inner edge the zone is essentially free of inclusions; but toward the outer edge it becomes crowded with angular, blocky xenolithic material and gradually passes into the shatter-zone of the country rock.

Immediately in beyond the contact zone is the marginal zone. Here the granite is coarser grained and grades inward to slightly coarser rock of the central zone, within roughly 5 or 6 hundred yards. The inclusions of the marginal zone differ strikingly from those of the contact zone. They are well rounded, highly elongate (tabular or lenticular), and show very perfect dimensional parallelism. The boundary between the contact zone and the marginal zone is marked by the sudden appearance of these inclusions in large size and number. Though the perfection of dimensional parallelism of inclusions remains essentially unchanged across the marginal zone, there is a gradual and noticable decrease in their size and abundance toward the center of the pluton.

Throughout the central zone inclusions uncommonly exceed 3 inches in length; only a few occur per square yard of outcrop surface; and preferred orientation is much less conspicuous.

The preferred orientation of these tabular and lenticular bodies establishes a planar structure for the rock. Throughout the central zone of the pluton this planar structure is essentially horizontal. In the marginal zone, however, the structure is inclined with dips ranging up to 35 degrees. Dip direction here is always radial, toward the interior of the pluton. The structure-pattern over the pluton, therefore, is considered (Chapman, 1969) to be saucer-shaped. It, thus, supports the idea that the hornblende granite formed in a floored magma chamber in the bottom of which feldspar (and perhaps hornblende and quartz) and xenoliths continued to accumulate, under direct gravitational control and influence of convection currents, to build upward a stratiform body of crystalline rock. This pattern of consolidation, from the bottom upward, appears representative of many of the other plutons.

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Itinerary

Mileage

0 Assembly point for trip is parking area at Acadia National Park Headquarters Building, 0.3 mile south of village of Hulls Cove (just off route Me. 3). Starting time 9:00 A.M. Park near the granite steps leading to the headquarters building. Try to arrive at least 30 minutes early to

permit a brief visit to the new Park Headquarters Building where complimentary copies of the park map may be obtained. As you climb the steps, note the large granite blocks used in the balustrade. This material, quarried at Stonington on Deer Isle, is much coarser than the granite of Mount Desert Island. Megacrysts of K-feldspar attain lengths of two inches and many show rapakivi structure.

Of the 22 stops described here, about 15 will be selected depending upon the weather, visibility, tide, size of group, and interests of participants. Stops of the most general interest are placed first for the benefit of those who may wish to leave the field party early.

Leave parking area; drive southeast on the "New Park Road" to Bar Harbor.

0.7 Stop 1. Paradise Hill Overlook. Park in area on left side of road. If visibility permits, the geology of the northern part of Frenchman Bay and surrounding region will be explained. Otherwise this stop will be ommitted.

1.2 New Duck Brook Bridge. Built of biotite granite from Hall Quarry in the town of Mount Desert. Just beyond the bridge are outcrops in the shatter-zone heavily stained with limonite.

2.0 Turn left to route Me. 3. On right is fresh cut in the shatter-zone showing somewhat rounded blocks of the Bar Harbor Series in granitic matrix.

2.8 STOP SIGN at route Me. 3. Cross route Me. 3 and proceed eastward into Bar Harbor village (on West Street).

3.5 Stop 2. Municipal pier. Park on the pier and walk to the outer end of the pier. Observe the structure of the Porcupine Islands to your front. Gently inclining beds of the Bar Harbor Series near the water level are overlain by thick cappings of a gabbro sill.

Now follow the "Shore Path" which starts near the pier and leads eastward, just above the sand beach and past the hotel. Examine the outcrop of mafic dike at the east end of the beach and, then beyond, the rocks along the shore (siltstone and sandstone of the Bar Harbor Series).

Return to the parking area. Drive on around the loop, leave the parking area, and start back down West Street.

3.6 Turn left up the hill on Main Street and drive southward through the village (on route Me. 3).

5.1 Junction of route Me. 3 and Schooner Head Road. Bear right on route Me. 3.

6.1	Turn right to Ocean Drive
6.2	Turn right to Ocean Drive
6.3	Turn right. You are now on a one-way road (Ocean Drive).
7.5	Stop 3. Champlain Mountain Overlook. Park in area at left side of road (elevation 253 feet). If the visibility permits, the geology of the southern part of Frenchman Bay will be explained. This location lies at the contact between the shatter-zone and the body of hornblende granite (coarse-grained granite of figure 1). At the south end of the road-cut is granite fairly typical of the contact zone of the Cadillac Mountain pluton. To the north the Bar Harbor Series is represented by huge blocks, with varicolored beds of metasiltstone and quartzite, enclosed or cut by veins and dikes of granite.
9.6	Stop 4. Sand Beach. A one-minute stop on road. Do not get out of cars. Look across the small pond to the ridge on the sky line. The light colored exposures on the ridge are granite, part of a small ring-dike within the shatter-zone.
11.8	Stop 5. SLOW!! Turn sharp right into Otter Point parking area and park. Walk out of the parking area by the east exit (same route that you entered). LOOK LEFT!! Then cross Ocean Drive and climb down onto the shore. Observe the huge mass (over 300 feet long) of varicolored quartzite and metasiltstone (Bar Harbor Series) with bedding dipping shoreward toward the body of hornblende granite. These beds were inclined downward along a ring-fracture zone as cauldron subsidence provided room for the influx of granitic magma. Walk westward along the shore, through a narrow shatter-zone, into the contact type of hornblende granite. Compare the granite in the outcrop here with that in the huge glacial boulders. Return to the parking area and drive out of the area by the west exit. LOOK LEFT!! Then turn right onto Ocean Drive.
12.7	Otter Creek causeway. To the north (right) from the causeway observe Cadillac Mountain (Elev. 1530 ft.) separated from Flying Squadron (Dorr) Mountain (elev. 1270 ft.) by a deep glaciated notch. Just ahead begins a long continuous road-cut through the marginal zone of the Cadillac Mountain pluton (hornblende granite).
14.4	Stop 6. Little Hunters Beach. Park off road at right. Walk across road, climb down to the cobble beach, and walk out along the west side of the cove to the high bare

ledges of breccia in the shatter-zone. Here beds of the Bar Harbor Series and large masses of dolerite (fine gabbro) have been involved in brecciation.

Return to the cars and continue driving along Ocean Drive.

14.7 Stop 7. Hunters Beach Head. A one-minute stop on left side of road. Do not get out of cars. Observe granite dikes cutting dolerite (fine gabbro) and beds of the Bar Harbor Series, all in the shatter zone.

17.5 ONE-WAY ROAD ENDS HERE!! Bear right onto Jordan Pond Road.

18.5 Road begins to climb west slope of Pemetic Mountain. Excellent road-cuts in coarse-grained granite of the central zone of the Cadillac Mountain pluton. Note jointing and inclined sheeting. To left is Jordon Pond in a U-shaped valley cut by glacial ice. Ahead are The Bubbles, two reshaped hills with steep slopes, plucked by the glacier, facing us. Near the crest of the nearest Bubble is a huge boulder, an erratic of very coarse granite carried from its source at least 20 miles to the northwest.

21.6 Stop 8. Eagle Lake overlook (elev. 485 ft.) A one-minute stop on left side of road. Do not get out of cars. Below is Eagle Lake. At its left end are The Bubbles exhibiting their asymmetrical outline. Beyond The Bubbles is Sargent Mountain (elev. 1373 ft.). On a clear day Blue Hill may be seen to the west, roughly 18 miles away.

22.3 Pass the Summit Road to Cadillac Mountain

22.9 Keep left at road junction

23.3 Eagle Lake Road overpass

23.4 Stop 9. Turn right on side road and then park on right about 50 yards ahead (just short of junction with route Me. 233. Walk to route Me. 233 just ahead and turn left. Proceed to the deep road-cut near the underpass. CAUTION!! STAY OFF THE ROAD! TRAFFIC USUALLY MOVES FAST HERE, AND VISIBILITY ALONG ROAD IS POOR. The hornblende granite here, in the marginal zone of the pluton, is a little coarser than that of the contact zone. Note the horizontal sheeting and steep jointing. Study the size, shape, arrangement, and distribution of xenoliths in the granite.

Return to the cars. Drive to route Me. 233.

23.5 Turn right onto route Me. 233.

24.8 Stop 10. LUNCH. Turn right into Eagle Lake parking area and park.

25.7 Stop 11. Park off road at right. Walk ahead to the low outcrops on right side of road. Compare the coarser, uniform texture of the hornblende granite here (central zone of the pluton) with that of the granite seen in the marginal and contact zones. Note the paucity of xenoliths. Just ahead the granite takes on an inequigranular texture (pseudoporphyritic). This rock is the recrystallized granite of figure 1.

26.5 Keep left on route Me. 233.

28.4 STOP SIGN. Junction with route Me. 198. Turn right. We now drive around the north end of Somes Sound, a flooded valley (fjord) cut deeply through the mountains to the south (left of road). Note medium-grained biotite granite (Somesville pluton) in several road-cuts.

29.9 STOP SIGN. Junction with route Me. 102. Turn left.

30.1 Enter Somesville, oldest settlement on Mount Desert Island. WARNING!! SPEED LIMIT STRICTLY ENFORCED.

30.8 Bear left at road junction toward Southwest Harbor.

32.1 Stop 12. At junction with Hall Quarry road. Park in area on right side of route Me. 102. Biotite granite of the Somesville pluton. This is the younger, finer-grained phase of the pluton which constitutes the youngest ring-dike of the area.

36.5 TRAFFIC LIGHT. Southwest Harbor village. Continue straight ahead.

37.1 Bear left at junction with route Me. 102A.

40.4 Turn left to Seawall Picnic Area on the shore.

40.6 Stop 13. Turn right and then park on the beach to the left of the road. Outcrop along shore represents weakly metamorphosed volcanic material (mostly ash and tuff) of the Cranberry Island Series.

Walk northward along shore a few hundred yards to observe mafic dikes cutting the metavolcanic rocks. If beach conditions are favorable, the contact between these rocks and a fine-grained sugary granite (miscellaneous granites of figure 1) may be observed. To the north, the granite.

speckled and mottled with hematite, carries abundant pods and veins of quartz and microcline (some amazonite).

Walk back to cars by the road and drive out to route Me. 102A.

40.8 Turn left onto route Me. 102A.

42.0 Turn right at Bass Harbor Head road.

44.3 Keep right at road junction.

44.6 Stop 14. Park on right side of road. Exposure of felsite of the Cranberry Island Series.

44.7 Road junction with route Me. 102. Turn left.

45.7 Keep right along the main road.

48.0 Stop 15. Junction with Dix Point road. Park along road. Exposure of fine-grained granite (granitic-granophyric phase of the Bays-of-Maine igneous complex).

50.0 Bridge over Seal Cove Brook. Turn left on side road just beyond bridge.

50.6 Stop 16. Park along road near shore of Seal Cove. If tidal conditions are favorable, rocks of the Ellsworth Schist may be studied. These somewhat pelitic schists show folded and crinkled bedding and foliation and abundant concordant quartz veins and pods.

Face the water and then turn left and walk along the shore to observe a granite-schist breccia, representing part of the shatter-zone formed during the intrusion of the granitic plutons.

Return to the cars and drive back to route Me. 102.

51.2 Turn left on route Me. 102.

51.3 Stop 17. Park along route Me. 102. Exposure of light colored quartz diorite of the gabbroic phase of the Bays-of-Maine complex.

54.4 Turn right on Long Pond Fire Road.

55.0 Stop 18. Park along road. About 250 feet west of highest point of fire road is side road to old quarry. Two phases (medium-grained and fine-grained) of the Somesville pluton may be observed here, and a small dike of the finer cuts the coarser rock.

Return to cars. Turn cars around at quarry road entrance and drive back to route Me. 102.

55.7 Turn right on route Me. 102.

56.2 Turn left at road junction.

56.5 Keep right at road junction.

59.2 Mount Desert - Bar Harbor townline.

60.6 Indian Point road junction, keep to left.

61.0 Stop 19. Park on side of road. Good exposure of the gabbroic phase of the Bays-of-Maine igneous complex.

62.4 STOP SIGN. Junction with routes Me. 102 and 198. Turn left.

64.4 STOP SIGN. Junction with route Me. 3. Continue straight ahead.

64.8 Stop 20. Turn right opposite information booth (before crossing bridge) into Thompson Island Picnic Area and park. Walk to the shore and examine the light greenish gray gneiss (Ellsworth Schist). Locally the rock is more properly a schist. Originally this material was probably well-bedded ash and tuff. Irregular, discordant quartz veins are numerous.

Drive back out of the parking area to the main road.

65.0 Turn left on route Me. 3.

65.4 Keep left on route Me. 3 at road junction.

68.6 Turn left onto Hadley Point Road and drive to the shore.

69.3 Stop 21. Hadley Point. Park on beach. At the water's edge, to your right front as you face the water, is a large outcrop of Ellsworth Schist. These somewhat pelitic rocks show marked foliation and crinkled layers. Concordant with the layers are thin, discontinuous veins and lenses of quartz.

Drive back up the hill to the main road.

70.0 Turn left on route Me. 3.

72.1 Turn left and follow sign to Sand Point.

72.8 Stop 22. Park off road on left near cabin area entrance. We have special permission from the property owners to walk down through the cabin area to the shore. Proceed to the shore, turn right at the beach, and follow the

shoreline to the high cliff with large caves at its base. CAUTION!! BE ON GUARD FOR FALLING ROCK! The cliff is held up by a microcrystalline rock (felsite) which in fresh exposure shows well-developed bedding. This material is part of the Bar Harbor Series, and more typical examples will be seen further along the beach.

Retrace your steps back to the cars.

73.7	Turn left onto route Me. 3 toward Hulls Cove.
75.2	Hulls Cove
75.5	Entrance to Acadia National Park. Turn right.
75.55	Turn right into Park Headquarters parking area. END OF TRIP.

STRATIGRAPHY AND STRUCTURE OF CENTRAL MAINE

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Introduction

The Merrimack Synclinorium broadens from Connecticut to the northeast, reaching a width of over 70 miles in central Maine. The stratigraphy and pre-metamorphic deformational history of the northwest limb were described in detail at the 1970 NEIGC Rangeley meeting (Moench and Boudette, 1970; Moench, 1970). The purpose of our trip is to examine the stratigraphy in the axial region and on the southeast limb of the synclinorium in order to demonstrate Silurian and Early Devonian regional lithofacies patterns for the now greatly compressed sedimentary trough. In view of this purpose, igneous and metamorphic petrology will only be briefly summarized here.

The trip includes stops in the Bangor, Dover-Foxcroft, Kingsbury, Guilford, and Skowhegan 15' quadrangles. All lie in what is generally (unfairly, in our opinion) referred to as the monotonous slate belt of central Maine. Metamorphic grade is for the most part low, ranging from garnet zone in the northwest to chlorite in the central and eastern part of the area. Exceptions are in the highly recrystallized inner aureoles of the four granitic plutons (Fig. 1) in which andalusite, sillimanite, and wollastonite are developed in pelitic and calc-silicate rocks.

Primary sedimentary structures are well preserved and facing indicators are abundant in most units. Graptolites have been found in 25 new localities, some of which will be visited on the trip. Ages suggested by W. B. N. Berry for these fauna provide firm Silurian ages for the lower part of the section, and confirm complex facies relationships for rocks of Llandovery through Ludlow age. Sedimentation was apparently conformable from at least Late Llandovery through Early Devonian times.

Stratigraphy

Central Maine was part of the Central Clastic Belt of Boucot (1968) from Llandovery through Ludlow times, and most units show the lithologies

and bedding characteristics of turbidites. Eight units of formation rank are recognized in the area of Figure 1. Fossil and structural evidence indicate the pre-Acadian deformation relationships represented schematically in Figure 2. In this paper, the metasedimentary rocks will be described by their protolith names without the prefix meta-. This is done to avoid awkward wording, and also because primary structures (and in some instances textures) are well preserved. It should be remembered that all rocks have been metamorphosed to at least the chlorite zone. Nomenclature is that currently used by the Maine Geological Survey and includes both formal and informal names.

Silurian System

Sangerville Formation: The Sangerville Formation underlies large parts of the Skowhegan, Kingsbury, Guilford, and Dover-Foxcroft quadrangles and is named from exposures on Route 23 between North Dexter and Sangerville (Guilford Quadrangel; Stops 5, 6, 7). It is composed for the most part of interbedded shales and coarser clastics (siltstone, sandstone) of graywacke composition, but limestone, granule conglomerate, and highly carbonaceous shale members have also been mapped. The Sangerville is highly heterogeneous; rapid lateral and vertical changes are common.

Graywacke and shale members (Ss): Well graded sandstone, siltstone, and shale with highly variable bedding thickness and proportion of lithologies make up most of the Sangerville (70%). Most exposures are of graded units 8 cm to 1 m thick in which sandstone and siltstone are dominant over shale (8:1-2:1). Thinner bedded, sometimes interlaminated units (1mm-4cm) are less common and contain equal proportions of sandstone and shale. Soft-sediment deformation (slump folds) is common at some horizons.

The shales weather dark gray to black, the coarser clastics buff, and color gradation indicative of increasing argillaceous content is often useful as a facing indicator. In the chlorite zone, shale and sandstone commonly contain a ferroan carbonate (ankerite, siderite) which weathers to hematite, yielding a characteristic light red-brown weathering rind. Clasts of quartz, feldspars, quartzite, chert, schist, slate, volcanic fragments ranging from sialic to mafic, and granitic hypabyssal rocks have been identified in Sangerville sandstones. These

EXPLANATION FOR FIGURE 1(following pages)

DEVONIAN

Dg, Di
Intrusive Rocks

Ds
Solon Fm.

SILURIAN

DSf
Fall Brook Fm.

DSv
Vassalboro Fm.

Sp
Parkman Hill Fm.

Ssg
Ssg
Ssl
Ssg
Ssl
Ssl
Ss
Sangerville Fm.
Ssg:conglomerate member
Ssl:limestone member

Scl
Sc
Canaan Fm.
Scl: limestone member

Swp
Swp
Swl
Swl
Sw
Waterville Fm.
Swl:limestone member
Swp:maroon and greem mem.

Skp
Skm
Skl
Sk
Unnamed Fm.
(Kenduskeag Fm)
Skp:maroon and green member
Skm:conglomerate
Skl:limestone

SCALE: 1:250,000

7
Trip Stop

23 2 (202)
Highway numbers

Railroad

Road

Route of Trip A-2(east half)

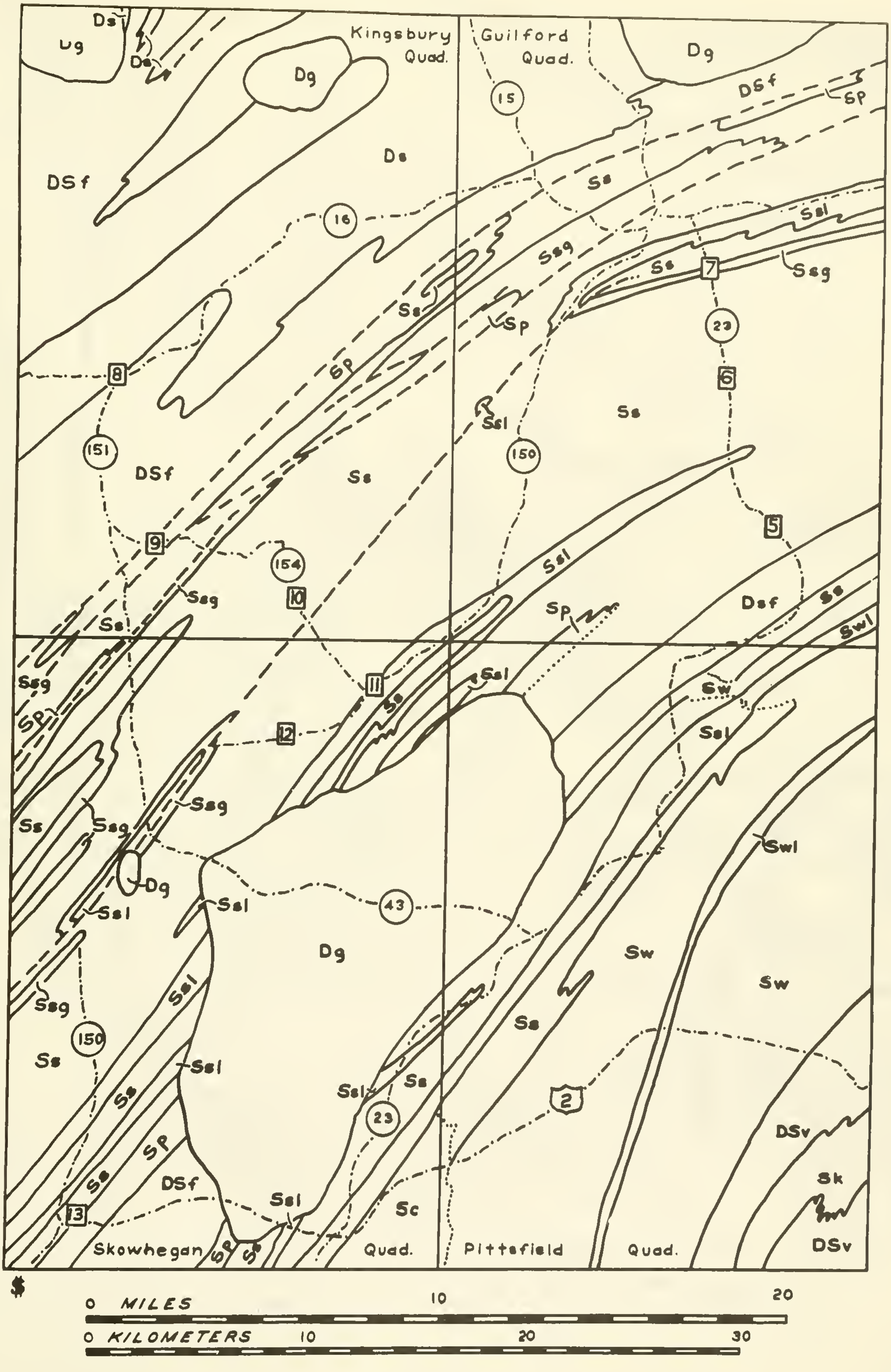
Route of Trip B-4(east half)

Interstate 95

Lithologic Contact

f
Faults

Facies contact


Kingsbury Quad.
Guilford Quad.
Skowhegan Quad.
Pittsfield Quad.
Dg
Ds
DSf
Ss
Ssg
Sp
Ssl
Sw
Swl
Sc
DSv
Sk
0 MILES 10 20
0 KILOMETERS 10 20 30

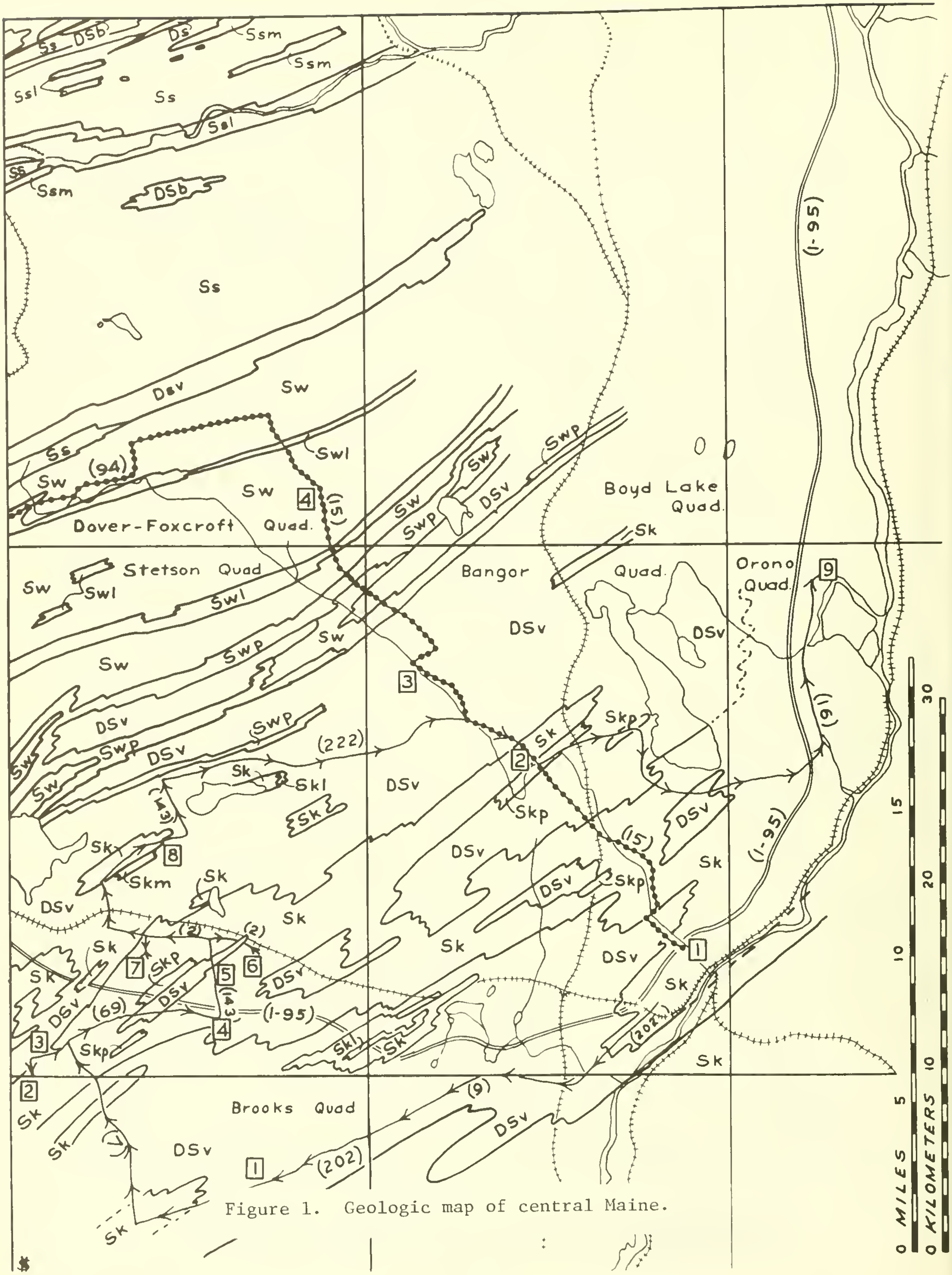


Figure 1. Geologic map of central Maine.

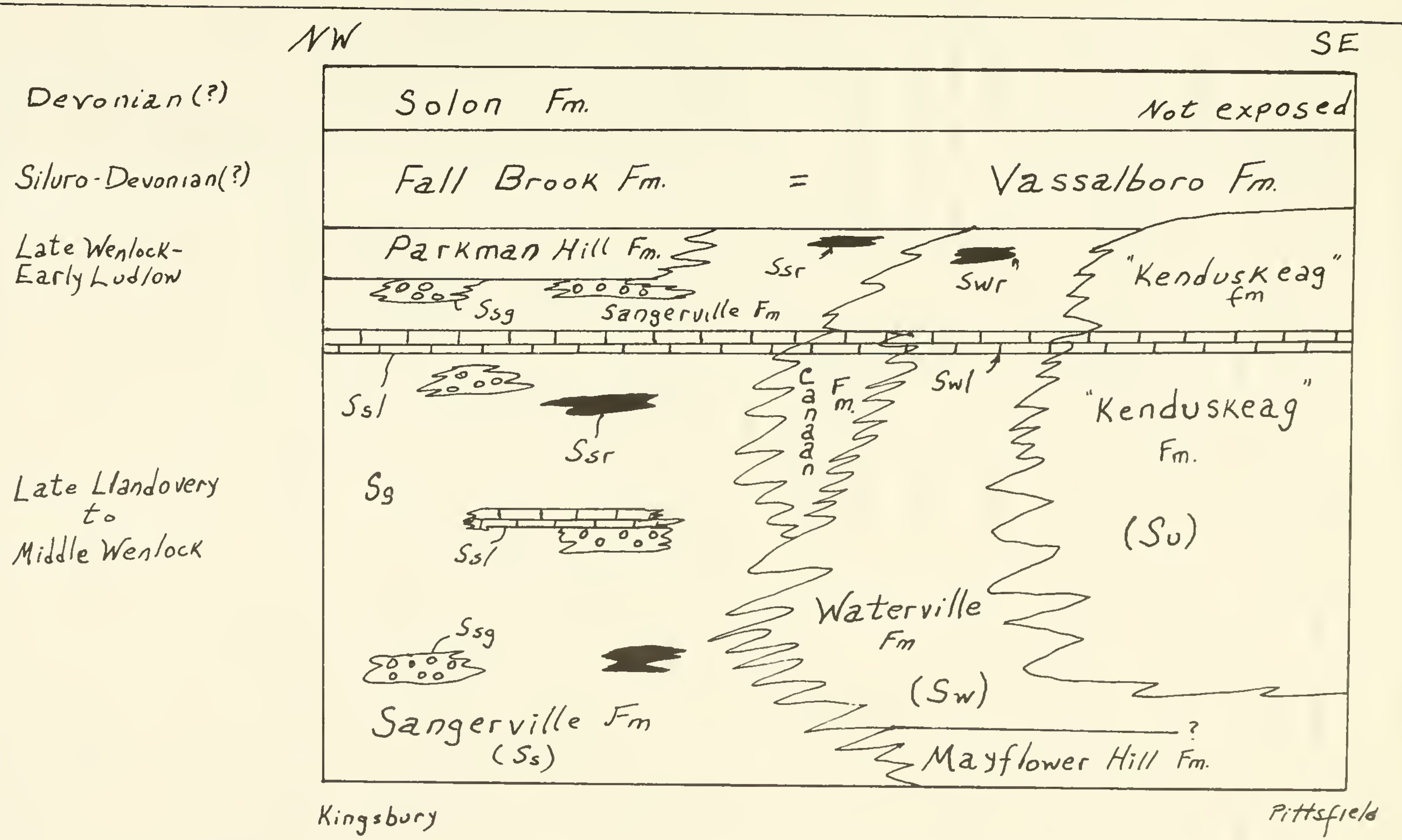


Figure 2- Inferred pre-Acadian Stratigraphic Relationships

are cemented by a calcareous and/or argillaceous matrix, often with large muscovite flakes. Sorting is poor.

Granule conglomerate member (Ssg): Whenever clast size exceeds 2 mm, rocks are mapped as Ssg. Two bedding styles are found but are not differentiated on Figure 1. Well graded units similar to those of Ss but with granule conglomerate at the base form continuous outcrop bands. Massive, non-graded, homogeneous granule conglomerate forms discontinuous lenses, particularly near the top of the formation. All clasts found in the sandstones have been identified in the conglomerates.

Ribbon limestone member (Ssl): Thinly interbedded silty micrite and non-calcareous siltstones, sandstones, and shales ("ribbon limestone") form several outcrop bands (Figure 1) but also occur as isolated lenses at different levels within the Sangerville. In contact aureoles, this member is recrystallized to calc-silicate granulite containing, in increasing order of metamorphic grade, clinozoisite, actinolite, diopside, grossularite garnet and idocrase, and wollastonite.

Carbonaceous shale member (Ssr): Highly carbonaceous shales, sometimes interlaminated with non-carbonaceous siltstones form lenses and discontinuous beds which could not be shown at the scale of Figure 1. The shale is often pyritiferous and rusty weathering, and contains all but one of the graptolite localities found in the Sangerville.

Because of intense folding and faulting, the internal stratigraphy of the Sangerville is incompletely understood at this time. All three minor members (Ssg, Ssl, Ssr) seem to occur at several different horizons. Thickness estimates are hazardous because of the intense deformation, and because the base of the Sangerville has not been recognized. An order of magnitude of thousands of feet is appropriate, but more exact figures cannot be given. Eight graptolite localities have been found in the Sangerville. In the type area, a range of Late Llandovery through Early Ludlow has been suggested by Berry. In the Kingsbury and Skowhegan quadrangles, however, the overlying Parkman Hill Formation, absent in the type area, contains graptolites as old as Upper Wenlock. In the western part of Figure 1, therefore, the Sangerville is dated as Late Llandovery through Middle Wenlock (Figure 2).

Canaan Formation (Sc): The Canaan Formation crops out sparsely in a small area surrounding the junction of the Skowhegan, Pittsfield,

Waterville, and Burnham quadrangles. It consists of thinly interbedded (1-5 cm) and interlaminated (1-5 mm) buff or light red-brown weathering siltstone of graywacke composition and dark gray to green shales. Silt/shale couplets are nearly always well graded, and delicate cross laminations are abundant in the upper parts of the siltstone. The siltstones are quite calcareous and contain ferroan carbonate. Ribbon limestone lenses and beds are abundant but granule conglomerate is absent. Siltstone:shale ratios range from 2:1 to 1:3.

The Canaan interfingers with the Sangerville Formation to the southwest and with the eastern facies of the Waterville Formation to the northeast. It represents a facies type intermediate between these formations; thinner bedded and finer grained than the Sangerville, slightly thicker bedded and more calcareous than the Waterville.

Waterville Formation (Sw): Rocks mapped as Sw in Figure 1 are on strike with those in the type area of the eastern facies of the Waterville Formation as redefined by Osberg (1968). The eastern facies of the Waterville Formation consists of thinly interbedded and interlaminated siltstones and shales which have a pinstriped appearance in large outcrops. A prominent ribbon limestone horizon was used by Osberg to delineate tight isoclinal folds in the formation, and a rusty weathering carbonaceous shale forms lenses near the limestone (Osberg, 1968; p 11) and also crops out discontinuously near the top of the unit (Osberg, personal communication). In addition, a distinctive sequence of green and maroon shales with interlaminated light gray siltstones is mapped near the top of the formation in the Pittsfield and Stetson quadrangles. Soft-sediment slumping is common.

Graptolites found in the Waterville Formation near Benton Station (Waterville Quadrangle) yielded a "Wenlock or Ludlow" age (Berry, _in_ Osberg, 1968; p 33). The _western_ facies of the Waterville Formation and the underlying Mayflower Hill Formation are considered to be equivalent to the Sangerville.

Unnamed Formation (Su): This unit, informally designated as the Kenduskeag Formation, crops out in the southern parts of the Pittsfield and Stetson quadrangles, and is characterized by extreme variability and the presence of rocks and bedding styles typical of the units described above. Granule conglomerate, ribbon limestone, carbonaceous

shale, and maroon and green shale are abundant but subordinate to graywacke/shale units with bedding similar to that of the Waterville and Vassalboro (see below) formation. These rock types are interlayered and interfinger irregularly; large outcrops may contain several different bedding styles.

Sedimentary breccias and chaotic zones are more abundant in this formation than in other units, and slump folding is most common here. Some of the coarser lithologies contain a high proportion of volcanoclastic debris, and some of the finer beds may be ash fall deposits.

Age and thickness are unknown at this time. The Kenduskeag is correlated with the Waterville and Sangerville formations, but may be somewhat younger in part (Figure 2).

Parkman Hill Formation (Sp): In the western part of Figure 1, the Sangerville is overlain by the distinctive rusty-weathering Parkman Hill Formation named by Pankiwskyj (in press) for exposures in the Anson Quadrangle. Over 60% of the formation consists of carbonaceous sulfidic shale interlaminated with non-carbonaceous, sparsely sulfidic siltstone and sandstone. Thick bedded (10-35 cm) sulfidic quartzose sandstones and non-sulfidic sandstones are also abundant, and are much less argillaceous than the comparable-sized clastics of the underlying Sangerville. Sulfidic granule conglomerate (somewhat more quartzose than the Sangerville varieties) and ribbon limestone are present, particularly near the top of the formation.

The Parkman Hill Formation thins to the east and interfingers with the Sangerville in the Guilford and Pittsfield quadrangles. In its stratigraphic position are discontinuous lenses of rusty-weathering carbonaceous shales (Waterville, Guilford, Dover-Foxcroft quadrangles) or the maroon and green shale of the Waterville Formation (Stetson, Pittsfield quadrangles). A maximum thickness of 300 m is suggested by Pankiwskyj (in press) for the western part of Figure 1. Well preserved graptolites yield Late Wenlock through Early Ludlow ages for the Parkman Hill Formation in the Kingsbury and Guilford quadrangles.

Siluro-Devonian(?)

Fall Brook Formation (DSf): The Fall Brook Formation was defined by Pankiwskyj (in press) as a series of generally thick-bedded calcareous sandstones with subordinate shales which lies above the Parkman Hill

Formation. The Fall Brook is divided here into three members, two of which will be seen on the trip. These members are not distinguished on Figure 1.

Thick bedded sandstone: Most of the Fall Brook consists of thick beds (20 cm - 2 m) of slightly calcareous sandstone and siltstone. Shale is an extremely minor component of this member. Outcrops appear massive but thin shale partings may be seen on fresh surfaces. Graded bedding is only rarely developed. Some highly calcareous layers weather rapidly and give a ribbed appearance to outcrops. At biotite grade and above, calc-silicate pods, stringers, and beds are common, and the formation has a light violet color due to finely disseminated biotite flakes. In unstained thin sections the sandstones seem to consist of well-sorted quartz grains in a biotite-calcite matrix. Muscovite is rare. Stained sections of the sandstone reveal up to 25% untwinned plagioclase feldspar in some specimens.

Thinly interbedded sandstone and shale: In several thin horizons within the Fall Brook, sandstones are thinly interbedded (up to 6 cm) with shale in graded sets. These sandstones are less calcareous than those of the massive member, and contain considerably more muscovite. The shale, at biotite grade and above, is typified by large (2 mm) poikiloblasts of biotite.

Transition member: The top of the Fall Brook consists of a series of well graded sandstones and shales in beds 10-30 cm thick. Feldspars and rock fragments up to 2 mm are found at the base of the graded sets, and scour-and-fill features are common. Calc-silicate beds and pods are abundant. Sandstone:shale ratios of 5:1 in these beds gradually diminish upwards and the beds thin. This gradation continues into the gray, dominantly shale base of the overlying Solon Formation.

Fossils have not been found in the Fall Brook Formation. It overlies the Early Ludlow fossils of the Parkman Hill and Sangerville formations and is definitely older than the Devonian(?) Solon Formation. A Siluro-Devonian age (Late Ludlow through Early Devonian) is assigned and a thickness of approximately 1,000 m is suggested.

Vassalboro Formation (DSv): Rocks very similar to those of the Fall Brook Formation and which lie above the Waterville and Sangerville formations in the Pittsfield and Stetson quadrangles are on strike with the type locality of the Vassalboro Formation as defined by Osberg (1968).

Bedding style, lithology, and thickness as described by Osberg are similar to those of the Fall Brook, but most Vassalboro outcrops are seen at lower metamorphic grade (chlorite) than the Fall Brook (biotite, garnet).

Devonian(?) System

Solon Formation (Ds): The youngest sedimentary rocks in the area of Figure 1 are the shales and rhythmically interbedded sandstone/shale graded sets of the Solon Formation, first named by Pankiwskyj (in press). The Solon is restricted to the northern part of the map area, presumably having been removed by erosion from the synclinal structures to the south.

The lowest part of the Solon is a dark gray silty shale with minor lighter gray weathering beds of siltstone which are commonly sheared and strung out along cleavage surfaces. These shales alternate with rhythmically interbedded light gray sandstones and shales above the base of the formation.

Poorly preserved brachiopods have been found in the Greenville Quadrangle in rocks mapped continuously with the Solon (Espenshade and Boudette, 1964; 1967). An Early Devonian age was suggested for these fauna and is assigned to the Solon.

Regional Correlation and Interpretation of Sedimentation

Correlation with the stratigraphy of the western limb of the Merrimack Synclinorium in Maine (Moench, 1970; Moench and Boudette, 1970) is shown in Table 1. The correlations are essentially those suggested by Osberg, Moench, and Warner (1968) but have the benefit of the recently discovered graptolite based ages.

Interpretation of regional lithofacies patterns shows that the filling of the original sedimentary trough was accomplished in two distinct stages. These are discussed here as the Llandovery-Early Ludlow and the post-Early Ludlow, although Ordovician metasediments almost certainly lie beneath the Sangerville.

Llandovery-Early Ludlow: The correlative Rangeley+Perry Mountain--Sangerville-Waterville formations are interpreted as proximal, intermediate, and distal turbidites respectively, derived from a geanticlinal terrain to the west and transported downslope to the east.

Table 1: Regional correlation in the Merrimack Synclinorium, Maine

	WNW		ESE
Devonian(?)	Seboomook Formation	Solon Formation	---
Siluro-Devonian	Madrid Formation	Fall Brook Formation	Vassalboro Formation
SILURIAN	Smalls Falls Formation	Parkman Hill Formation	
	Perry Mountain Formation	Sangerville Formation	Waterville Formation
	Rangeley Formation		Mayflower Hill Formation

Comparison of thickness estimates and grain sizes of equivalent units shows that formations thin and grain size decreases toward the east. What little faunal evidence is available supports the paleoslope inference, and indicates a bathyal to abyssal environment (Griffin, 1973). Bedding thickens and grain size increases again in the Kenduskeag to the east. An eastern source area, presumably the Silurian coastal volcanic terrain of Gates (1967) is inferred for this formation. The eastern third of Figure 1 lay at or near the axis of the original sedimentary trough and shows the interfingering of sediments derived from both east and west. Reduction of the western source area led to the deposition of the generally fine grained Smalls Falls and Parkman Hill formations in an euxinic environment.

Post Early Ludlow: The environments of deposition described above and characterized by rapid lateral lithologic variations were abruptly ended by deposition of an apparently homogeneous thick blanket of sandstone (Madrid-Fall Brook-Vassalboro), followed by a thick turbidite sequence (Seboomook-Solon) in which direction of sediment transport is difficult to interpret. This interruption may have been due to uplift associated with the Ludlovian Salinic Disturbance (Boucot, 1968).

Structure

Outcrops in central Maine are characterized by northeast strikes and vertical and near-vertical dips. The rocks have been subjected to intense polyphase deformation--folding of four generations and faulting--and intrusion by granitic plutons. Folds seen in outcrop may be attributed to one of the four stages of tectonic deformation or to soft-sediment slumping.

Slump Folds

Slump folds are tight isoclinal folds of small scale (2.5 cm - 1.5 m wavelength). Larger scale slump folds may be present but cannot be demonstrated due to limitation of outcrop size. Slumps occur as right laterally asymmetric anticline-syncline sets restricted to a narrow stratigraphic horizon and bounded by "undisturbed" beds. The folded horizon is commonly separated from confining beds by a sedimentary decollément at the base and a welded contact at the top. The rocks deformed plastically, presumably while still partially saturated with water, and large folded fragments in sedimentary breccia are common. The slump folds have been rotated into the vertical position by the first major tectonic deformation, and have been cut by cleavages associated with the tectonic folds. Typical features of slump folds are shown in Figure 1 of Trip A2, this volume.

Tectonic Folds

Four stages of tectonic folding (F_1, F_2, F_3, F_4) are recorded in central Maine, but the first was the most important.

F_1 folding was responsible for the major fold axes and tight isoclinal minor folds in central Maine. Despite the vertical nature of most beds, it is thought that the large scale folds (10-20 km wavelength) are relatively open structures. Small scale folds include those with wavelengths of 120-370 m and .3-2 m. The 120-370 m wavelength folds can often be traced across several quadrangles. This fact, the parallelism of fold limbs over 10's of kilometers, and the generally gentle plunges of minor fold axes indicate that the regional structures are gently (doubly) plunging. An axial plane cleavage is associated with F_1 folds, and the larger clasts have their longest axes aligned parallel to this cleavage. Response to F_1 deformation is dependent on rock type. Ribbon limestones and thinly bedded silt/shale

sets are extremely tightly folded with very sharp hinges while thick bedded sandstones and granule conglomerates have blunt hinges and are somewhat more open. Shearing has removed the hinge areas of many F_1 folds so that opposing limbs are juxtaposed along the shear surface. In many such instances, the west-facing limb of the fold has been shortened. The F_1 folds are truncated by Middle Devonian granitic plutons and are considered to be the result of the main phase of the Acadian Orogeny.

The map pattern shown in Figure 1 is dominated by a sinusoidal change in regional trend from N30E in the west, through N70-80E in the central part, to N50E in the east. This is the only mappable expression of F_2 folding. Most F_2 folds are small scale features (5-20 cm wavelength). In the eastern part of Figure 1, two conjugate folds are attributed to F_2 deformation. Folds with vertical axial surfaces (F_{2v}) are most abundant; folds with horizontal axial surfaces (F_{2h}) are subordinate, possibly because most outcrops are two-dimensional pavements. F_2 folds are best developed in thin-bedded rocks (ribbon limestones, Waterville-type pinstriped rocks). A strongly penetrative axial plane fracture cleavage (N10E-N10W) associated with F_{2v} is well developed, however, in thicker bedded rocks. This cleavage is nearly parallel to F_1 cleavage in the west, but is nearly perpendicular to it in the central and eastern parts of the map area. Plunges of F_{2v} folds are moderate to steep (40-80°). In the Stetson Quadrangle, F_2 folding is clearly post-metamorphic. It is thus post-major Acadian deformation but may be due to some later phase of that orogeny.

Small scale open warping of F_2 and F_1 folds is visible in some large outcrops and is designated as F_3. Axial planes of these folds trend N60-70W and plunges are steep. Axial plane cleavage is not well developed, but the most common joint set parallels the axial trend.

A set of east-west trending, generally left-lateral kinks which can be observed in the thinner bedded lithologies in the eastern part of Figure 1 is the last product of folding in the area and is designated as F_4. F_4 folding has not been observed in the western part of the area.

Faults

Intense shearing occurs in the northwestern part of Figure 1. High angle faults are mapped by the presence of gouge, silicified breccias, mylonite, outcrops consisting entirely of vein quartz, and by sharp discontinuities in map trends of stratigraphic units. The faulting was a brittle phenomenon; sandstones are badly shattered and injected with veins of quartz and calcite. Displacement and age of the faulting are not fully understood at this time. Movement in most cases appears to have been dip-slip with the northwest block downdropped. In some cases, slickensides suggest strike-slip movement and map patterns suggest right-lateral movement.

F_1 folds are truncated by the faults, but the relationships between the shearing and the later fold episodes are unclear. Intrusion of the Porcupine Mountain apophysis of the Hartland Pluton (Skowhegan Quadrangle) seems to post-date faulting; in this case, faulting should be pre-contact metamorphism. F_2 folds, however, are post-metamorphic and do not seem to be disturbed by the faulting.

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Itinerary

Mileage	
	Meet in parking lot behind the Mammoth Mart on Harlow St. (immediately west of U.S. Post Office). Trip will assemble near the Morse Covered Bridge.
0	Upon leaving the parking lot turn left and drive west on Harlow St.
0.2	Bridge over Kenduskeag Stream.
0.3	Pull over to right and park on shoulder of road.
	<u>Stop 1</u>. "Kenduskeag formation". Sequence of massive fine-grained quartzite beds one to two feet thick interbedded with thin alternating layers of phyllite and coarse metasiltstone. The latter are ¼ to ½ inch thick. Sedimentary breccia, possible slump folds, tight isoclinal folds (F_1), small-scale horizontal folds (F_2), and minor shears (S_5) can be observed.
	Return to vehicles and proceed northwest on Harlow St.
0.5	Bear left (south) at "Y" onto Fourteenth Street.
0.9	Stop sign. Turn right (NW) onto Ohio Street.
1.1	I-95 Overpass. Continue straight.
2.4	Stop sign. Turn right (NE) onto Griffin Road.
2.5	Bridge over Kenduskeag Stream. Outcrop of "Kenduskeag formation" to right (S).
2.7	Blinking yellow traffic light. Turn left (N) onto Kenduskeag Road.
4.3	Stop sign. Turn left (W) onto Route 15 North (Broadway St.)
5.7	Bridge over Kenduskeag Stream at Six Mile Falls. Outcrops of "Kenduskeag formation" to right in stream.
6.5	Bangor & Aroostook Railroad overpass. Outcrop at left (S) is "Kenduskeag formation", intruded by a plagioclase granulite dike.
9.4	Bridge over Kenduskeag Stream. Park on right shoulder by trailer camp. The outcrop to be viewed is under the bridge.

Stop 2. Maroon and green phyllite member of "Kenduskeag formation".

Sequences of maroon and green phyllite and coarse metasiltstone with beds ½" to 1" thick occur in both the "Kenduskeag formation" and the Waterville formation.

Graded beds and left-lateral kink bands (F_4) are present. Trace fossils have been found in this lithology although as yet none have been observed at this outcrop.

Return to vehicles and continue north on Route 15.

10.2 Outcrop of "Kenduskeag formation" on left (SW).

12.3 Outcrop of "Kenduskeag formation" on left in Kenduskeag Stream.

12.5 Town of Kenduskeag. Gas available.

13.4 Road runs along top of esker here.

13.8 Bear left (west) onto side road at "Covered Bridge" sign.

15.4 Turn sharp left (south).

15.7 Robyville Covered Bridge (1876).

Drive through bridge and park on shoulder of road. The outcrops to be viewed are on the south bank of the stream, west of the bridge.

Stop 3. Vassalboro formation.
Massive fine-grained quartzite in beds 2 to 3 feet thick. Some of the beds are graded, and are phyllitic at the top. Tops are to the south.

Turn around and drive north on the dirt road.

16.2 Five-way intersection. Take second road to right and proceed east.

17.3 Intersection with Route 15. Turn left (north). Esker.

18.2 Corinth Knoll Picnic Area.

19.2 Outcrops of Waterville formation on both sides of road.

22.4 Town of East Corinth. Gas available. Continue on Route 15.

24.0 Small outcrops of "ribbon lime" in fields on either side of road.

24.1 Intersection with Route 11. Continue on Route 15. View of Charleston Ridge on horizon ahead. The ridge is underlain by Vassalboro formation.

25.9 Park on shoulder of road at top of hill. View of Charleston Air Force Base ahead.

Stop 4. Waterville formation.
Thin alternating phyllite and metasilstone layers 1/8" to 1/4" thick.

Isoclinal folds (F_1) and axial plane cleavage (S_1) are present. Some of the isoclinal folds are probably of slump origin and there may be a sedimentary decollement present (see Griffin and Lindsley-Griffin, Trip A-2). Both horizontal and vertical folds (F_2) and their associated fracture cleavages (S_2) may also be seen.

Return to vehicles and continue north on Route 15.

27.3 Outcrop of Waterville formation on right (east).

27.5 Town of West Charleston. Turn left (west) onto side road.

29.4 Outcrop of Waterville formation on right (north).

29.9 Road crosses an asker.

30.2 Outcrop of Waterville formation on left (south).

31.6 Outcrops of Waterville formation on both sides of road.

32.3 Intersection. Turn left (south).

32.3 Outcrops of Waterville formation.

33.3 Intersection. Town of Garland. Turn right (west) onto Route 94 West.

33.4 View of Garland Pond to left (south). "Ribbon lime" member of Waterville formation exposed on shore.

34.3 Outcrop of Waterville formation on right (north).

39.3 Just beyond bend in Route 94 bear right onto dirt road by Puffers Pond.

39.6 Outcrop of ribbon lime on right (east).

39.9-40.4 Small outcrops of Waterville formation on both sides of road.

40.4 Contact with "Sangerville formation".

40.6 Intersection with paved road. Turn right (east).

40.9 At "Y" bear right. Park off road.

Optional Stop A. Waterville-"Sangerville" contact.
Outcrops of "Sangerville" and Waterville formations separated by an 80' covered interval.

Northern outcrop: "Sangerville formation". Well-graded, fine-grained, quartzite beds 6" to 1' thick with black phyllite tops 2" to 3" thick. Tops are to the north.

Southern outcrop: Waterville formation. Thin-bedded alternating phyllite and metasiltstone layers ½" thick.

Turn around at Dexter Dump and return eastward along road.

41.5 Intersection with dirt road. Continue straight. Remain on Main Street to stop light.

42.1 Traffic light in Dexter. Turn right (north) onto Route 7.

42.4 Turn left onto Route 23 (Dam Street). Cross Maine Central Railroad tracks and bear right on Route 23.

42.8 Outcrop of "Sangerville formation" on left (west).

42.9 Lake Wassookeag. Continue straight.

44.0 "Y" intersection. Remain on Route 23.

44.9 Outcrops of "Sangerville formation".

45.2 Outcrops of "Sangerville formation".

45.8 Park off road in parking lot.

Stop 5. "Sangerville formation" and rusty-weathering phyllite member.

Fine-grained calcareous quartzite, with graded beds, cross-laminae, and possible sedimentary breccia.

Northeast-plunging disharmonic isoclinal fold (F_1), and axial plane cleavage (S_1) are present.

Ninety feet to the south is a small roadcut outcrop on the east side of Route 23 of black carbonaceous phyllite with well-developed axial plane cleavage (S_1). Weathered surfaces are rusty-colored due to weathering of pyrite.

Graptolites collected from this locality have been identified by W.B.N. Berry (pers. comm. to Griffin, 1967, 1970). The fauna consists of:

Cyrtograptus ? sp.
Monograptus praedubius ?
Monograptus sp. (of the *M. dubius* group)
Monograptus sp. (in the *M. priodon*--*M. flemingii* group--possibly *M. flemingii*.

Berry (pers. comm., 1967) states that the age is "In the span of Late Llandovery--Wenlock. If the *M.* sp. in the *M. priodon*--*M. flemingii* group is indeed *M. flemingii* then this collection would indicate a Wenlock age."

Continue north on Route 23.

46.1 Outcrops of "Sangerville formation".

46.7	Bridge over stream at North Dexter. Outcrop of "ribbon bon lime" in stream.
47.4-47.6	Outcrops of "Sangerville formation" along road.
47.8	Outcrop of "Sangerville formation". Bedding is near-horizontal here because it is at the crest of a major fold (F_1). Outcrop also affected by second folding (F_2).
48.9-49.3	Outcrops of "Sangerville formation" along road. Good graded bedding visible.
49.5	Park on shoulder of road just before the top of the hill.

Stop 6. "Sangerville formation".
Fine-grained calcareous quartzite, well-graded, with black phyllitic tops. Beds are 6" to 3' thick, phyllite layers are 2" to 10" thick. Tops are to the south.

This is a typical outcrop of the "Sangerville formation". Axial plane cleavage (S_1) and a fracture cleavage (S_2) are present.

Return to vehicles and proceed north on Route 23.

49.8-50.4	Outcrops of "Sangerville formation".
50.6	Pleasant Acres Picnic Area.

Continue north on Route 23.

51.4-51.7	Outcrops of "Sangerville formation".
51.8	View of Piscataquis River Valley; Manhanock Pond to left.

Just beyond Sangerville town line sign, park on shoulder of road.

Stop 7. Polymictic granule conglomerate member of the "Sangerville formation".
Consists of thick beds (here 10'), sometimes graded, of quartz, feldspar, and phyllite chips.

Continue north on Route 23.

52.1	Outcrop of polymictic conglomerate on left (west).
53.0	Town of Sangerville. Outcrops of "ribbon lime", coarse quartzite, and calcareous quartzite in Carlton Stream north of bridge.

Proceed north through town on Route 23.

53.4	Piscataquis River.
53.5	Bear left at "Y" onto Route 16. For the next 2.2 miles, outcrops are "Sangerville formation".

54.8 Turn left, remaining on Route 16, and cross the Piscataquis River.

54.9 Turn right, drive west on Route 16.

55.7 Senior Citizens Home; the flagpole in the yard is set in an outcrop of polymictic conglomerate.

58.5 Town of Abbot Village. Cross the Piscataquis River and remain on Route 16. Outcrop of "Fall Brook formation" visible in river.

58.6 Turn left on Route 16 to Bingham.

61.9 Side road on left, immediately followed by outcrops of "Solon formation".

62.5 Bridge over Thorn Brook, a tributary of Kingsbury Stream.

65.0 Outcrop of "Solon formation" on right.

70.3 Kingsbury Center. Remain on Route 16.

70.5 Kingsbury Pond picnic area.

71.1 Mayfield town line.

72.9 Turn left on Route 151 and park.

<u>Stop 8</u>. Contact between Fall Brook and Solon formations

Start traverse on north side of Route 16 at culvert and walk west on Route 16. Metamorphic zone is <u>garnet</u>, highest grade seen on the trip.

Calcareous biotite-metasandstone with rare pelitic partings representative of typical Fall Brook metasandstone passes upward (to the west) into well graded metasandstone, metasiltstone, and phyllite with garnetiferous calc-silicate granulite layers and pods. Abundant facing indicators (graded beds, scour and fill, load phenomena) show that tops are toward the northwest. Bedding thins and pelite becomes more abundant to the northwest.

Cross to south side of road at break in outcrop and continue west.

First outcrops on south side of Route 16 are typical gray phyllites of the Solon Formation, with minor light gray metasiltstone laminae and disrupted laminae. The mineral assemblage in the phyllites here is muscovite-biotite-garnet-chloritoid-quartz-plagioclase. Chlorite is present as a retrograde product from chloritoid and garnet.

Cleavages associated with both F_1 and F_{2v} are well developed in the Solon phyllites, less well displayed in the Fall Brook, and cleavage/cleavage lineations are prominent.

Return to cars and continue south on Route 151

74.6 Poor outcrop of Fall Brook sandstones on right

76.8 Brighton center; turn left (east) onto Route 154 toward Wellington

77.3 to 78.0 Small outcrops of Fall Brook to left

78.8 Cross outlet of Trout Pond

78.9 Stop 9: Parkman Hill Formation; park on right shoulder.

Several Parkman Hill lithologies are visible here. Poorly preserved fragments of monograptids from this locality have been assigned a Silurian (?) age by W.B.N. Berry. Rock types include:

Highly carbonaceous pelite and intercalated sulfidic metasiltstone. Pyrite stringers are parallel to cleavage. Rusty weathering illustrated here is typical of the formation.

Non-sulfidic metasandstone and finely interlaminated pelite and metasiltstone are interlayered with the "rusties" and are clearly distinguished by a lack of the characterstic weathering.

Minor sulfidic quartzose metasandstone occurs in beds 8cm thick.

Continue east on Route 154. If time permits, log from 79.6 to 82.1 will be followed. If not, continue on Route 154.

79.6 Turn right at fork onto dirt road.

79.7 Pass outcrops of medium bedded Sangerville metasandstones.

79.8 Pass rusty-weathering Parkman Hill carbonaceous pelites containing well-preserved monograptids indicative of Early Ludlow-Late Wenlock age.

80.2 Cross excellent pavement outcrops of Sangerville turbidites.

80.5 Entrance to field on right; small fault slice of Parkman Hill Fm.

81.5 Turn left onto dirt road. Pavement outcrops are of thinly interlaminated pelite and siltstone of the Sangerville Fm.

82.1 Turn right onto Route 154.

83.0 Pass Sangerville metasandstones in roadside and stream outcrops to right.

83.2 Wellington; turn right (south) toward Harmony.

83.9 Pass chlorite zone Sangerville metasandstone and pelite on right.

85.4 Stop 10: Small pavement outcrop of Sangerville Formation metasandstone and pelite in the chlorite zone. Sandstones and pelite here are highly ankeritic; weathering of ankerite to hematite produces typical reddish brown spotted appearance of Sangerville in the chlorite zone.

Deformation in the sandstones appears to be soft-sediment slumping modified by tectonic shearing.

Continue south on Route 154.

88.2 Junction with Route 150. Continue across 150 bearing slightly to the right.

88.5 Bear right into Harmony center; park in front of Barber Salon on the right.

Stop 11: Walk north into Higgins Brook and climb to extensive exposures of Sangerville ribbon limestone at the Harmony dam.

NO HAMMERS PLEASE

Three generations of folds (F_1, F_2, F_3) are clearly visible, with marker beds of feldspathic metasandstone and boudinaged quartz veins being extremely helpful in outlining the folds.

Extremely tight isoclinal folds with vertical axial surfaces (F_1) trend N40E, essentially parallel to the spillway of the dam. F_1 cleavage is well developed in the micrite layers of the limestone.

F_1 isoclinal folds are refolded by N10E trending right-handed asymmetric minor folds visible across the entire outcrop but best displayed by a feldspathic metasandstone bed near the dam. F_2 cleavage is well developed in both micrite and sandstone beds. Both F_1 and F_2 cleavages are filled with quartz and calcite.

F_3 is represented by open warping best illustrated by the quartz vein close to the dam. Joints at the outcrop are parallel to the N65W axial trend of the warps.

Continue west on road, across Higgins Brook and uphill.

88.9 Turn left (west) onto Route 150

90.4 Pass outcrops of Sangerville sandstone and pelite on right.

91.7 Pass Athens town line and park.

Stop 12: Badly fractured Sangerville sandstone and pelite is on strike with Stop 10, but is at biotite grade here. Ankerite is replaced by prograde pseudomorphs composed of biotite, muscovite, and quartz. Graded bedding is clearly visible.

F_1 axial plane foliation in the pelites is clearly out by by the F_2 cleavage, best illustrated at the top of the outcrop.

Continue west on Route 150

92.0	Pass graptolite-bearing (possibly Wenlock) highly carbonaceous pyritiferous pelite in the Sangerville associated with massive granule conglomerate to west.
94.3	Pass Sangerville sandstone and pelite on left
96.2	Junction with Route 151; bear left into Athens
96.4	In Athens center at junction with Route 43. Bear Right on Route 150 towards Skowhegan.
96.5	Cross Wesserunsett Stream. Excellent Sangerville ribbon limestone to left and about 25-100 yards upstream. Sangerville sandstones just upstream of bridge, carbonaceous pelite and granule conglomerate downstream.
96.5	Route 150 swings to right; hill at 11 o'clock is Porcupine Mountain apophysis of the Hartland Pluton.
97.3	Calc-silicate hornfels in Porcupine Mountain aureole.
98.4	Junction with Route 43; continue south on Route 150
98.9	Sangerville siltstones and pelite cut by small andesitic dike under the powerline at left.
99.9	Sangerville sandstones and pelites in fields to both sides of road
105.2	Sangerville ribbon limestone on hillside to left.
107.5	Bear left at intersection north of Skowhegan center.
108.3	Junction (unmarked) with Route 2 near center of Skowhegan. Turn left (east) onto Route 2.
109.0	Park opposite gas station and walk on path to Great Eddy of the Kennebec River. Beware of the poison ivy.

Stop 13: Contact between Sangerville and Parkman Hill formations.

The uppermost Sangerville rocks here are medium to thin bedded metasandstones and phyllites in well graded and strongly laminated units reminiscent of the uppermost rocks of the Perry Mountain Formation at Smalls Falls (Moench, and Boudette, 1970). Large prograde pseudomorphs after ankerite are prominent in the pelite. Calcareous pods (not yet calc-silicate) are abundant near the top of the Sangerville.

Parkman rocks here include carbonaceous pelite and thin beds of sulfidic metasiltstone. Sulfied (pyrite and pyrrhotite) occurs as blebs elongated parallel to F_1 cleavage in the pelite and siltstone. Massive sulfidic and non-sulfidic quartzose sandstones of the Parkman Hill Formation are exposed 0.1 mile to the east on Route 2.

Trip ends here. To rejoin Route 95 south, make a U-turn and drive back to Skowhegan. Drive past the intersection with Route 150 at the post office and follow signs to Route 201 south. Take 201 south until entrance to Interstate 95.

To return to Orono, continue east on Route 2 to Newport, then follow signs to entrance to Interstate 95 NORTH to Orono.

ECONOMIC DEPOSITS AT BLUE HILL

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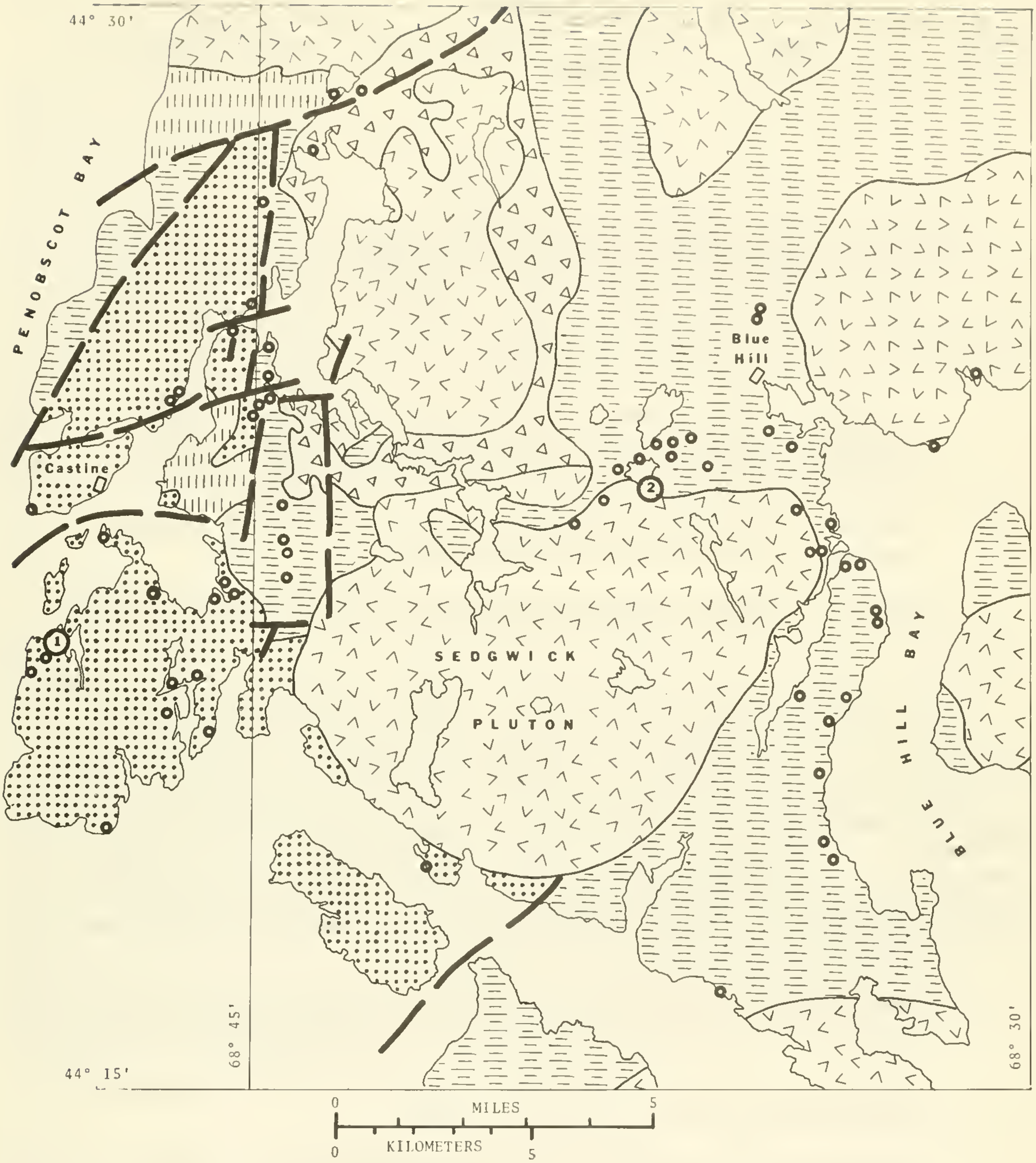
Introduction

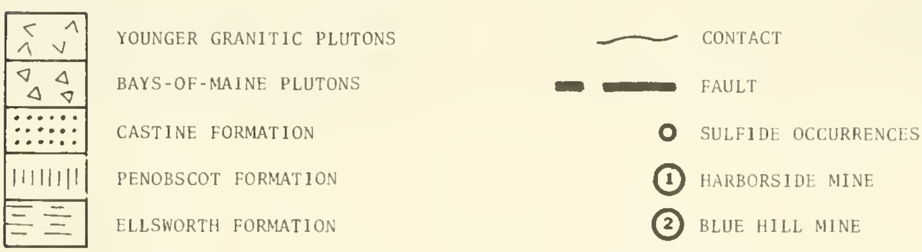
The Penobscot peninsula, which lies between Penobscot Bay and Blue Hill Bay, is underlain by schists, volcanic rocks and intrusive rocks (see fig. 1). The Ellsworth Formation, which is dominant in the eastern part of the area, is a heterogeneous sequence of interlayered pyroclastic rocks, flows and clastic sedimentary rocks which have been highly deformed and metamorphosed to chlorite or biotite grade. At some localities original bedding features have been preserved, but more commonly the primary characteristics have been obliterated. Typically the rocks of the Ellsworth Formation contain abundant oriented chlorite or biotite resulting in a strong foliation. Colors range from light gray or greenish gray to almost black, depending on the relative abundance of quartz and feldspar as compared to the abundant chlorite and biotite. Segregations of quartz-rich zones are commonly present parallel to the foliation resulting in banded structures. Interbedded with the highly contorted schistose rocks are quartzite units as much as 300 feet thick which have reacted competently to the deformation.

The Castine Formation which dominates the western part of the area is composed of volcanic rocks including pillow basalts, felsic to intermediate tuffs and rhyolitic breccias, and less commonly slates and phyllites. The great variation in lithology has made stratigraphic correlation, even on a local scale very difficult, and as a result the total thickness and regional stratigraphy have been difficult to assess. The rocks have been subjected to chlorite grade metamorphism, but characteristically the original textures, especially of the coarser pyroclastic rocks have been preserved. Brookins et al. (1973) have established the age of the Castine volcanic rocks as 390 $\pm$ 5 million years.

The intrusive rocks of the area include gabbro and diorite of the Bays-of-Maine complex (Emmons, 1910; Chapman, 1962; Cheney, 1969), and the younger plutons most recently described by Chapman (1968) and Wones (1974). The younger plutons which show a close spatial relationship to many sulfide occurrences are characteristically granite or quartz-monzonite which vary in texture from coarsely porphyritic to fine-grained equigranular. They exhibit sharp discordant contacts not only with the Ellsworth and Castine Formations, but also intrude the rocks of the Bays-of-Maine complex. Commonly the intruded rocks have been brecciated, and the dislocated blocks have been engulfed and assimilated to varying degrees by the felsic plutons. Where the younger plutons invade the Ellsworth Formation, extensive metamorphic aureoles have resulted. The most common effect of metamorphism is the formation of hornfels containing biotite and cordierite as the characteristic minerals. Faul (1963) has determined the younger plutons to be early Late



EXPLANATION



Geology Modified from Brookins et al. (1973), Chapman (1968) and Cheney (1969)

FIGURE 1. GENERALIZED GEOLOGIC MAP OF THE BLUE HILL - CASTINE AREA

Devonian in age. Brookins (1968) determined the Sedgwick pluton, one of the younger intrusives, to be 395 $\pm$ 15 million years, and also points out that it intrudes the Castine Formation.

Sulfide Occurrences

Those sulfide occurrences located in the western part of the map area are concentrated in the Castine volcanic rocks; those in the Blue Hill area are concentrated at or near the contact between the Sedgwick pluton and the Ellsworth Formation (see fig. 1). This suggests several possibilities regarding the origin of the sulfide deposits, two of which will be pursued briefly.

1. - A single period of sulfide mineralization which post-dated the younger plutons and which produced hydrothermal replacement deposits over a large area in receptive host rocks.

2. - Two periods of sulfide mineralization, each of which affected a somewhat restricted geologic environment.

At this point, some of the characteristics of the ore deposits should be examined to provide evidence of their origin.

The Harborside deposit, which was the site of an open pit mine operated by the Callahan Mining Company from 1968 to 1972, is the largest known occurrence of those associated with the Castine Formation. Sphalerite and chalcopyrite are the dominant ore minerals, accompanied by abundant pyrite and lesser amounts of pyrrhotite. The host rocks are relatively undeformed, slightly metamorphosed (chlorite grade) or hydrothermally altered (chloritic) pyroclastics. At that deposit there are many textural and mineralogical characteristics which indicate that fluids of very low viscosity permeated the host rock and allowed deposition of sulfide minerals by a process commonly referred to as hydrothermal replacement. Some of the more common and, at the same time, spectacular developments are:

1. - Presence of unmineralized rock fragments which differ in composition from the mineralized matrix of the pyroclastic host rocks.

2. - Selective sulfide mineralization of certain minerals and grains in the matrix of the pyroclastic rocks.

3. - Preservation of pyroclastic and metamorphic textures and structures within the sulfidized zones.

4. - Presence of sulfide metacrysts transecting original pyroclastic textures.

5. - Presence of doubly terminated sulfide crystals in pyroclastic rocks.

6. - Presence of gradational sulfide mineralization fronts.

No one of these features by itself is conclusive evidence of hydrothermal replacement, but the combination of them in a single deposit is compelling evidence that the sulfide mineralization took place after the pyroclastic rocks were at least partially solidified. Most of the characteristics indicate that the rock was completely solid and had already been subjected to its slight metamorphism (or alteration) prior to the sulfide mineralization. These conclusions do not preclude the possibility that the ore-depositing solutions were derived from the same source as the volcanic material and represent a late stage of the volcanogenic cycle.

The largest and most intensively studied of the sulfide occurrences near the contact of the Sedgwick pluton and the Ellsworth Formation is the Blue Hill ore deposit, which is being mined by Kerramerican, Inc. by underground methods. Sphalerite and chalcopyrite are the dominant ore minerals; galena is rare and of no economic importance, and pyrite and pyrrhotite are common gangue minerals. The most important host rocks are slightly deformed quartzite units of the Ellsworth Formation. Locally within the mine area thin (25 to 50 feet) off-shoots of the Sedgwick pluton contain significant sulfide mineralization.

There are several metamorphic and hydrothermal minerals which are important in determining the sequence of events in the Blue Hill deposit. Contact metamorphism of the chlorite rich metasedimentary rocks has produced biotite-cordierite rocks whose textural variations allow them to be classed as either hornfels, schist or gneiss. These contact metamorphosed rocks have in turn been replaced by sulfide minerals; replacement textures are well shown by biotite-sulfide relationships. In addition, dravite (brown tourmaline) is present in veins and irregular masses which cross-cut both the contact metamorphosed Ellsworth rocks and the Sedgwick pluton. The tourmaline in turn is replaced by sulfide minerals, most commonly chalcopyrite and pyrrhotite. These relationships clearly indicate that the sulfides were transported in the ionic state and are of post-intrusive age. The close spatial relationship between the sulfides and the pluton indicates the probability of either a collinear or cognate genetic association between the intrusive and sulfide mineralization. Ching (1942) has outlined a sequence of events indicating a collinear relationship, but the sequential associations outlined above for the Blue Hill deposit are equally indicative of a cognate relationship as described by Burnham (1967).

History of the Blue Hill Mine

The discovery of copper mineralization in 1876 on the north shore of Second Pond led to the development of the Douglas mine and smelter which produced 2,000,000 pounds of copper from 1880 to 1884. Following an extended period of dormancy, the mine was reactivated for a few months during

World War I, but could not survive the low market price of copper. Another period of quiescence preceded an attempt by the U.S. Bureau of Mines to delineate an eastward extension of mineralization in 1948 by drilling seven exploratory holes. Lack of sulfide intercepts discouraged further work at that time.

Recent exploration began in 1957 when Texas Gulf Sulphur instigated a drilling program which resulted in the discovery of significant copper and zinc mineralization directly beneath Second Pond and slightly to the southwest. Black Hawk Mining Company, a subsidiary of Denison Mines, Ltd. then entered the scene and pursued those discoveries with additional drilling through the early 1960's. Encouraged by the results of their drilling project, Black Hawk expanded their efforts by sinking a shaft in 1964 and 1965. The three-compartment shaft reached a depth of 698 feet, with development levels at 380, 480 and 580 feet. Black Hawk completed approximately 10,000 feet of lateral development on those three levels and in addition drilled 31,750 feet of core from the underground workings. The company experienced difficulty in holding experienced miners with the project and as a result also found it difficult to keep the development with the ore. Early in 1967 Black Hawk Mining Co. suspended operations and permitted the underground workings to become flooded.

In 1970 Keradamex, Inc., a wholly-owned American subsidiary of Kerr Addison Mines, Ltd., entered into an option agreement with Black Hawk Mining Co. for the development of and production from the Black Hawk mine. After completing a drilling program to test a geologic theory of continuity of ore mineralization, Kerradamex exercised the option in mid-1971. By agreement, Kerramerican, Inc. (another subsidiory of Kerr Addison established to mine this property) was to acquire a 60% interest in the property after producing 500 tons of ore per day by September 1, 1973. Construction and pre-development began in July 1971 and in September of that year a 15% decline ramp was started in order to provide trackless access from the surface to the ore horizons. Once the ore zones were reached, a trackless pilot and slash technique was used for mining. Two rubber-tired jumbos were used for drilling, and broken ore was loaded into dump trucks for transportation to the crusher or stockpile at the surface. Kerramerican met their production objective as outlined in the agreement with Black Hawk Mining Co., and now are completely responsible for the mining and milling operation.

Mine Geology

The Second Pond mineralized area is underlain by the Ellsworth Formation which has been intruded and contact metamorphosed by the Sedgwick pluton. That portion of the Ellsworth Formation which is present in the mine area can be subdivided into four lithologic units which are described briefly in Table 1.

TABLE 1. Subdivisions of the Ellsworth Formation in the Blue Hill Mine.

Unit	Thickness	Description
Allen quartzite	greater than 300 feet	Massive to banded, brownish-gray to purplish-gray biotite-cordierite quartzite; variation in color and foliation dependent on amounts of dark brown biotite and blue-gray cordierite each of which may range in content from near zero to 40%; gradational into underlying schist with increase in biotite content. (Originally this quartzite was considered two units, Allen quartzite and Robbins quartzite, but is now treated by Kerramerican as a single unit).
Biotite schist	100 to 500 feet	Foliated, crenulated brownish-black to purplish-black biotite schist; biotite is dominant (up to 85%), blue-gray cordierite porphyroblasts are common in some areas, absent in others, quartzite content is variable and inversely related to the biotite content.
Pond quartzite	150 to 300 feet	Massive, gray quartzite with minor biotite and sericite; generally lacks prominent foliation and recognizable bedding; host for most of the known ore deposits.
Banded quartzite	150 to 250 feet	Upper portion dominated by alternating brownish-black biotite rich layers and dark green chloritic quartzites; quartzite dominates the lower portion and resembles the Allen quartzite, with the exception of the chlorite content.

The most common effect of metamorphism by the intrusive is the transformation of chlorite to biotite. In certain stratigraphic horizons where the mineralogy is suitable and where heat has been sufficient, cordierite porphyroblasts have developed and have produced a rock with a gneissic texture. Biotite and cordierite are not characteristic of the Ellsworth Formation except near the intrusive contacts.

In addition to the mineralogical change, a hybrid breccia zone has been formed irregularly at the contact. It is composed of partially assimilated

blocks of Ellsworth Formation surrounded by granitic material. Both the mineralogical and intrusive breccia effects form zones as much as several hundred feet thick.

Although the Sedgwick pluton has a relatively narrow range in mineralogical composition (granite to quartz monzonite), the texture varies from fine-grained to coarse-grained and porphyritic, and the color varies from light gray to blue-gray, green and tan-orange. The color variations are primarily due to feldspar hue; although there is a wide range in the biotite content, its effect is to lighten or darken the colors. The pluton tends to be concordant and sill-like, generally located above the Pond quartzite. Within the mine area the thickness of the pluton is variable, attaining a maximum of 200 feet and increasing in thickness toward the south. There are many localities near the contact where the pluton has been partially to almost completely replaced by sulfide minerals. To complicate matters, there are also a few localities where unmineralized apophyses of the pluton cut across sulfide zones in the Ellsworth Formation.

A significant thickness of dark green, medium-grained gabbro is associated with the Sedgwick pluton southwest of the mine area. Within the mine, however, only minor (10 feet or less) gabbro or diabase dikes occur. All observed gabbro occurrences in the mine are pre-ore in age, and presumably part of the Bays-of-Maine igneous complex.

Ore deposits

The ore mineralization is related to the axis of an open syncline in the Ellsworth Formation striking N 30°W and plunging 30°SE. Mine development and geologic studies have disclosed a significant wrinkling in the axial plane of the syncline which may have resulted from drag folding or superimposition of a second generation fold. The Ellsworth Formation and banded ore deposits commonly contain folds with amplitudes of 10 to 20 feet. Several folds of large magnitude have been observed and interpreted to be distorted drag folds parallel to the major fold. Interpretation of surface drilling indicates a flattening of the major syncline down plunge, but data are sparse and details of the structural trends are not well known.

Several faults which are locally significant have been recognized on the mine property. The Mammoth fault was first detected by Black Hawk Mining Company as a result of surface drilling and has been further defined by Kerramerican drilling. Early in 1974 the fault was crossed by the Mammoth haulage drive, substantiating its projected position. New movement on the fault is estimated to be 100 to 200 feet in a reverse direction.

The Carleton fault was outlined by Kerramerican geologists on the basis of drill hole intersections which show planar continuity. The throw on the Carleton fault is estimated to be 100 feet or less in a reverse direction.

The Sedgwick and West faults, which are cross-faults between the Carleton and Mammoth faults, were first detected in drill hole intersections. The West fault has been substantiated by mine development, and the Sedgwick fault is now interpreted to be a locally sheared monoclinal fold.

The Dam fault is the only one of significance recognized in an ore body. It is present in the Second Pond "A" zinc zone where it has a displacement of about 30 feet, and is accompanied by marked thickening of massive sulfides on the hanging wall. The combined accumulation of sphalerite, pyrrhotite, pyrite and chalcopyrite approaching 30 feet is unique in the mine.

The most important ore mineralization at the Blue Hill mine consists of massive sphalerite occurring in at least four strata-bound horizons which are nearly continuous over most of the mine area. The horizon which currently accounts for the greatest production is the Second Pond "A" zinc zone which occurs near the upper contact of the Pond quartzite. The mineralization has local folds and disruptions of sufficient magnitude that in some areas the miners have difficulty staying on the ore. Massive sphalerite of the "A" zone ranges in thickness from less than an inch to over twenty feet, and averages about two and a half feet. The massive sphalerite of the "A" zone contains as much as 3% chalcopyrite and 25% combined pyrite and pyrrhotite; commonly the zone is dominated by gangue sulfides. Disseminated and veinlet sulfide minerals occur throughout the mine area, most commonly near the massive sulfide horizons. Sphalerite is typically coarse-grained, dark brown to almost black; rarely paler shades of brown and fine grained textures occur. Three additional zinc horizons similar to the "A" zone have been discovered by diamond drilling, but ore bodies have not yet been delineated.

Chalcopyrite mineralization is widespread in the mine area, commonly occurring as replacement at intergranular boundaries in quartzitic units. It also occurs in fracture fillings and by replacement in all rock units including the Sedgwick pluton. In spite of the widespread occurrence of chalcopyrite, only two zones are large enough and sufficiently mineralized to be considered orebodies. The Lower Second Pond (L.S.P.) orebody lies directly beneath the "A" zinc zone, at or near the base of the Pond quartzite. Thickness of the L.S.P. orebody ranges from a few feet to 30 feet. The Mammoth area, southwest of Second Pond contains two zones similar to the L.S.P. orebody, but on the basis of current information, only a portion of one of the zones in the lower-middle Pond quartzite can be considered an orebody.

Hydrothermal alteration of host rocks is common but not consistent throughout the mine. A systematic study concerning alteration has not yet been carried out, and at this time the spatial and paragenetic relationships between alteration products and sulfide minerals are not well known. Sericite, chlorite, biotite, amphiboles, tourmaline and green feldspar have been reported as alteration products, but analytical confirmation has been attempted for only a few specimens.

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Itinerary

Discussion of mine geology and tour of underground workings will take place at the Kerramerican Blue Hill mine (about 1½ hour drive from Orono). To reach the Blue Hill mine follow Route 15 from Bangor to Brewer to Bucksport and Blue Hill. Directions from Blue Hill village to the mine are as follows:

Mileage	
0	Blue Hill Post Office at intersection of Route 15 with Routes 172 and 176. Follow 15, 172 and 176 southwest out of town.
0.6	Intersection. Continue straight ahead on 15 and 176.
1.8	Turn left on gravel road at sign "Kerramerican Blue Hill Venture".
2.1	Mill tailings pond on left.
2.2	Second Pond on right.
2.4	Mine gate. Proceed through gate and bear right. Continue to right of concrete block building with mine headframe. Park in lot overlooking Second Pond. Discussion begins at 10 A.M.

THE CONCENTRICALLY ZONED TUNK LAKE PLUTON: DEVONIAN MELTING-ANOMALY ACTIVITY?

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Introduction

The post-tectonic Tunk Lake granite pluton (Fig. 1) is exposed over a circular area of 180 km^2 in southeastern Maine. It intrudes the Ordovician (?) Ellsworth Schist, the middle Paleozoic Bays-of-Maine gabbroic-granitic complex, and Middle Devonian biotite granite and quartz monzonite (Karner, 1968). A preliminary K-Ar date (Karner, in preparation) is late Middle Devonian (357 ± 10 MY).

The purpose of this paper is to summarize briefly the geology of the pluton following closely the work reported in Karner (1968), and Karner and Helgesen (1970) and also to discuss the pluton's possible relationships to the Maine Coastal plutons, the White Mountain plutonic-volcanic series, the tectonic history of the northern Appalachians, and to possible New England melting-anomaly activity as discussed in the above references, Karner and Bertram (1972) and Karner (1973, and in preparation).

Rock Types

The outermost rocks of the pluton are magnetite-aegirine augite granites containing sodium-rich microperthite. These grade inward to hornblende and biotite granites and biotite quartz monzonite. Six rock types, distributed in concentric zones, can be distinguished by texture, quartz and feldspar contents, and common ferromagnesian minerals (Table 1). Their outcrop areas are shown on Figure 1 except for type I which occurs at the margin of the pluton. In adjacent zones, rock types are gradational so that contacts between zones are arbitrary and have been placed according to characteristics given in Table 1.

Type I: Magnetite-Aegirine Augite Granite

Type I occurs in the contact zone of the pluton. Table 1 describes its texture and major mineral content. Magnetite and aegirine augite occur as subhedral to euhedral grains 1 to 2 mm long. Microperthite occurs as subhedral tabular grains from 1/2 to 3/4 cm long. Some phases of the granite contain up to about 15 percent oligoclase as discrete grains from 1/4 to 1/2 cm long. Quartz occurs as anhedral grains from 1/4 to 1/2 cm in diameter. Euhedral zircon, sphene, apatite, and allanite-epidote are common accessories. Fayalite is present in some specimens. The contact zone, about 30 meters wide at the east margin

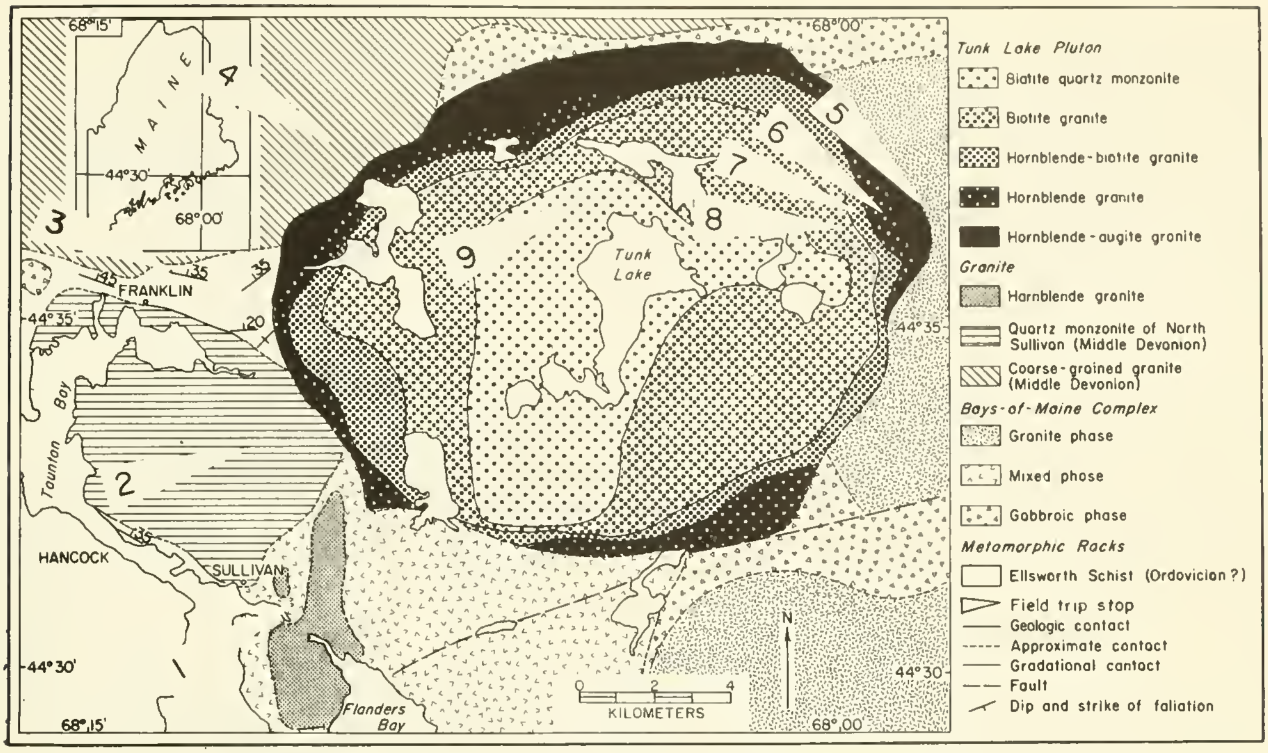

Figure 1. Geologic map of the Tunk Lake granite pluton (Karner, 1968).

Zone	Rock Type	Texture	Areal Extent	Quartz	Oligo-clase	Perthite	Albite in Perthite	Total Mafics + Acces-sories	Mag-netite	Aegir-ine Augite	Horn-blende	Biotite
I..........	Magnetite-aegirine augite granite	Medium-grained hypidiomorphic granular	....	30–40	<10	55–65	40–45	2–4	1	<1	<1	<1
II.........	Hornblende-aegirine augite granite	Medium-coarse-grained hypidio-morphic gran-ular	8.7	20–25	< 5	60–70	45–50	6–10	1	1–2	3–5	<1
III.......	Hornblende granite	Medium-coarse-grained hypidio-morphic gran-ular	10.2	25–30	< 5	55–65	40–50	4–7	1	<1	2–4	1
IV.........	Hornblende-biotite gran-ite	Medium-coarse-grained hypidio-morphic gran-ular	13.2	25–35	5–15	45–55	30–40	4–6	<1	....	1–2	1–3
V..........	Biotite granite	Medium-coarse-grained hypidio-morphic gran-ular to porphy-ritic	40.4	30–35	15–25	40–50	25–35	3–5	<1	....	<1	2–4
VI.........	Biotite quartz monzonite	Medium-grained porphyritic to allotriomorphic granular	27.5	30–40	20–30	30–40	15–25	2–4	<1	....	......	2–4
Average..	....	....	....	31	18	46	31	5	0.6	0.1	1.1	2.5

Note.—Areal extent expressed as percentage of total outcrop area. Mineral abundances expressed as volume percentages. Albite in perthite expressed as percentage of total perthite. Average weighted according to areal extent of zone. Modified from Karner (1968).

Table 1. Characteristics of rock types of the Tunk Lake granite pluton (Karner and Helgesen, 1970).

of the pluton, is interpreted as a chill zone for the following reasons: (1) Position: It occurs only at the outer margin of the pluton where it grades inward into zone II. (2) Texture: The rock is medium-grained. Types within the pluton are medium-coarse-and coarse-grained. (3) Mineralogy: The rock contains minerals which crystallized early in the formation of the pluton. Within the pluton these early minerals have recrystallized during slower cooling. (4) Composition: The bulk composition of the chill zone is close to the average for the pluton at the exposed level (Table 1).

Type II: Hornblende-Aegirine Augite Granite

Calcic hornblende, probably containing significant sodium and ferric iron, occurs as subhedral to euhedral grains from 2 to 3 mm long with cores of aegirine augite. Perthite occurs as subhedral grains from 1 to 2 cm long, and quartz, as anhedral grains from 1/4 to 1/2 cm in diameter. Riebeckite and a reddish-colored biotite (probably astrophyllite) are common in small amounts. Accessory minerals are similar to those of the chill zone.

Type III: Hornblende Granite

Type III resembles type II but contains very little aegirine augite. Albite intergrowths in perthite are irregular, forming patch perthites. A small amount of quartz is intergrown graphically-granophyrically with perthite. Accessory minerals include magnetite, zircon, apatite, allanite-epidote, and some fluorite. Small amounts of biotite and sphene are usually present.

Type IV: Hornblende-Biotite Granite

In type IV hornblende is present as subhedral grains which characteristically are partly rimmed and replaced along cleavage cracks by biotite. Some perthite grains have rims of oligoclase, and intergrown quartz is common. Most oligoclase occurs as subhedral grains from 1/4 to 1/2 cm long. Accessory minerals are similar to those of the hornblende granite.

Type V: Biotite Granite

In type V texture is variable. Fine- to medium-grained aggregates of perthite, oligoclase, quartz, and mafic minerals occur between coarser grains. Aggregates may be narrow zones between a few perthite grains, or they may be well developed and may completely isolate larger grains of quartz, feldspar, and mafic minerals. Accordingly, the texture may by hypidiomorphic-granular, seriate-porphyritic, sub-porphyritic, or porphyritic. Seriate textures are the most common. Patch perthite contains less exsolved albite than in outer zones and commonly has oligoclase rims (An_{21}) forming typical rapakivi texture. Oligoclase also occurs as discrete grains (An_{13}) up to one centimeter long. Quartz, graphically-granophyrically intergrown with perthite, is common. Much biotite is partly chloritized. Magnetite, sphene, and apatite are common accessories. Zircon and allanite-epidote are less abundant and less well developed than in rocks closer to the margins of the pluton. Fluorite is found locally.

Type VI: Biotite Quartz Monzonite

Textures of type VI are gradational with those of type V but medium-grained phases with texture similar to the groundmass of porphyritic types are most common. Mafic minerals include biotite, chloritized biotite, chlorite, and magnetite. Patch perthite contains very little intergrown albite. Oligoclase (An_{11}) has clear albite rims (An_3) and altered cores richer in calcium (An_{20}). Apatite and subhedral-anhedral zircon and sphene are common accessory minerals. Less common are fluorite, molybdenite, and allanite-epidote.

Finer-Grained Phases

In each of the six types, finer-grained variants are found within rock of normal texture. These phases are especially common toward the center of the pluton occurring as fine- to medium-grained irregular or sheetlike masses from several inches to many feet in width. In the biotite granite and quartz monzonite many of these masses may occur within an outcrop. There are gradations among aplites dikes, less regular fine-grained dike-like or sill-like bodies, and irregular fine-grained masses. Mineralogy generally resembles that of normal types closer to the center of the pluton. In the biotite granite and quartz monzonite these phases are often granophyric and miarolitic.

Form and Structures

In the pluton a resistant margin forms an outer rim of high hills, except to the south where shearing has weakened the granite. At the present level of erosion the pluton transects metamorphic and igneous rock masses of different age (Fig. 1). To the south contacts are unexposed or are in areas where the rocks are sheared and relations to the adjacent Bays-of-Maine complex are unclear. The dip of the contact surface of the pluton was not found in outcrops observed. In areas of contact with gabbro-diorite, a breccia occurs which contains subangular xenoliths up to at least several feet across. Outward from this zone the rocks and actual contact are covered. Characteristic topography in the contact zone is a steep outward slope at the outermost rock exposures. An abrupt change in magnetic properties at the edges of the pluton suggests steep contacts. Oriented fine-grained mafic xenoliths, slab-like and usually a few to several inches long, strike parallel to the contact of the granite and the country rock in the northwestern northeastern, and eastern margins of the pluton and dip inward at angles of about 30° to 40°, suggesting that the contacts of the zones may dip inward and that the outer contact may also dip inward (Karner and Connors, 1971). Rocks of the pluton are generally massive. Parallelism of tabular perthite grains causes a faint layered appearance in a few specimens from the contact phase. Perthite grain alignment around some inclusions suggests that the parallelism of feldspar in these rocks is a primary flow structure. Oriented xenoliths may also indicate magmatic flow.

Major joint trends for the pluton are E-W, WNW, NE, and ENE. Most

joints are vertical or steep. Subhorizontal sheeting is common. Aplite dikes occur throughout the pluton but are most common in the central part. Major trends are NW and NNE. Dikes, often a few inches wide and steeply dipping, consist of abundant quartz and sodium-poor perthite, subordinate plagioclase, and small amounts of biotite, chlorite, and opaque minerals. Often the dikes have central openings almost completely filled by inward-projecting feldspar and quartz crystals up to several centimeters across. Quartz veins, an inch or so wide, and zones of quartz veinlets sometimes occur in the central zones. Major trends are E-W, WNW, and ENE. Mafic dikes are rare. One, 2-4 m wide is exposed at location U 6.5, V 22.1 (Fig. 2). It contains abundant xenoliths of granite and with several adjacent narrower dikes trends east-west and is approximately vertical. Sulfide mineralization is present on faces of a WNW joint set at the location mentioned above. Molybdenite occurs in the central zones in altered granite and in quartz and aplitic veins particularly on Catherine Mountain (Trefethen and Miller, 1947, Young, 1963).

Compositional Variation

Compositional variation in the pluton has been studied primarily by modal analysis of thin sections. Some chemical work has been done and extensive major and trace element studies of both bulk composition and mineral phases have been undertaken. Bedrock exposures are usually found on the sides and tops of hills 15 m or more in height. Exposures present in low areas indicate that they are often underlain by rock which is more strongly fractured or more mafic than that of the higher well-exposed parts of the pluton. The low areas are generally those marked unsampled on Figure 2. Specimens from 140 field stations (Fig. 2) were chosen for study of compositional variation. A series of closely spaced specimen locations was chosen along the E-W Tunk Lake section. The mineralogical composition of the pluton varies systematically from the margins inward (Karner, 1968; Table 1; Fig. 2). In Figure 2 values for the thin marginal zone I are not shown. Modal variation may be summarized as follows:

(1) The quartz content decreases from an average of 34 percent in zone I to 22 percent in zone II, and then increases to 35 percent in the core, zone VI.

(2) The perthite content decreases from 57 and 65 percent in zones I and II to 35 percent in the core zone VI, while its albite content decreases from 45 percent to 21 percent. Oligoclase varies antipathetically with perthite and increases from about 5 percent in zones I and II to 27 percent in zone VI.

(3) The dominant mafic mineral varies according to the series magnetite-aegirine augite-hornblende-biotite. Total mafics and accessories average 3 percent in zone I, increase to 8 percent in zone II, and decrease to 4 percent in the core, zone VI.

(4) Modal data vary regularly from the margin of the pluton inward. Contours generally parallel the outer contact of the pluton and the zone boundaries as recognized in the field. Total mafics and accessories show the greatest deviation from regular variation.

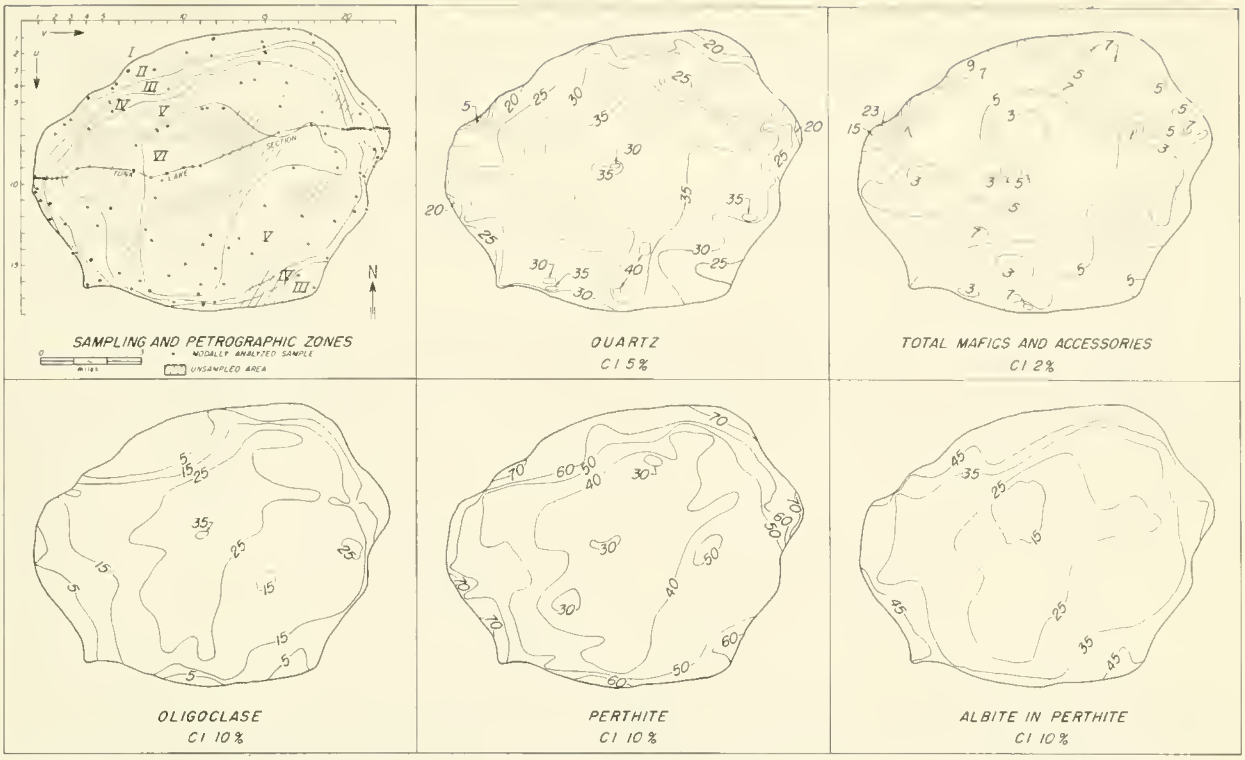

Figure 2. Modal variation (vol. %) in the Tunk Lake pluton (Karner, 1968).

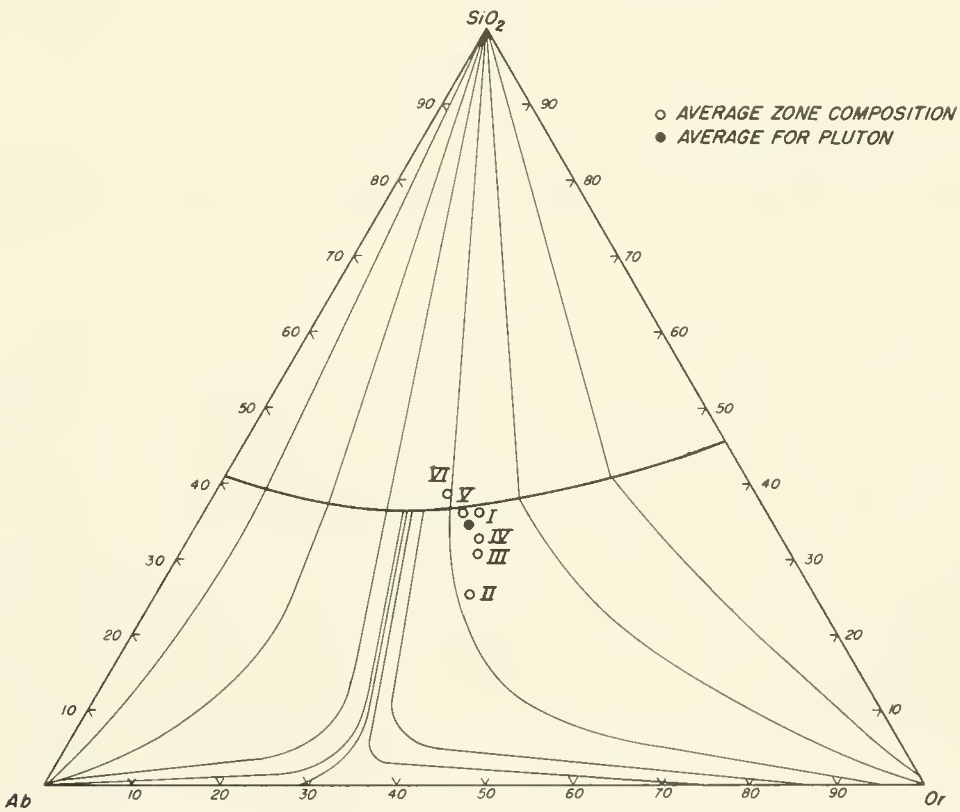

Figure 3. Average compositions for the zones of the Tunk Lake pluton plotted on a diagram showing isobaric fractionation curves for a water-vapor pressure of 1000 kg/cm^2 in the system $NaAlSi_3O_8$-$KAlSi_3O_8$-SiO_2-H_2O (Tuttle and Bowen, 1958).

(5) Zone I closely approximates the average for the pluton in content of quartz, feldspar and total mafics and accessories.

Wet and calculated (from modes) chemical analyses may be summarized as follows:

(1) SiO_2 decreases from an average of 76 percent in zone I to 70 percent in zone II and then increases to 76 percent in the core, closely paralleling the variation of quartz.

(2) Al_2O_3, $Na_2O + K_2O$, $FeO + Fe_2O_3$, and CaO increase from values close to the average for the pluton in zone I to highs in zone II and then decrease to the core corresponding to the variation of total feldspar and total mafics and accessories.

(3) The ratio of Na_2O to K_2O is highest in zone II.

Emplacement and Crystallization

Emplacement of the pluton at an assumed depth from 2-5 km occurred by piecemeal stoping along marginal zones of ring fracturing and by cauldron subsidence or stoping of large blocks in the central zones. The magma at the time of intrusion is assumed to have had a composition, indicated by a calculated average for the pluton, approximately equal to the biotite granite of zone V, a volatile content from 2 or 3 percent, and a temperature of 800° C.

Using these assumptions the crystallization history of a magma with a composition similar to that of the average for the rocks of the pluton can be summarized. Experimental data from the 1000 kg/cm^2 isobaric equilibrium diagram for the system Ab-Or-SiO_2-H_2O (Tuttle and Bowen, 1958) show phase relations at approximately the conditions of beginning of crystallization of the pluton (Fig. 3).

Crystallization began in the outer portions of the pluton with precipitation of magnetite and aegirine augite followed by sodium-rich alkali feldspar and quartz. Exsolution of feldspar was followed by the partial replacement of primary mafic minerals by hornblende and biotite. During crystallization, silica-rich aqueous fluids migrated from the margins to the center of the pluton changing the bulk composition of successive internal zones of the pluton (Fig. 3). The marginal rocks were left quartz-poor, and feldspar- and mafic-rich. The core rocks became quartz-rich. In the central zones additional water lowered crystallization temperatures allowing oligoclase, hornblende, and biotite to form as primary minerals and also facilitated late-stage recrystallization. Recrystallization fluxed by hydrous fluids accounts for fine-grained intergranular material in the central granites and also for sheets, lenses, and irregular bodies of fine-grained rock throughout the pluton. Fluids may have moved inward and upward in response to a pressure gradient caused by eruption at the surface.

Assimilation of mafic rock is a possible cause of the enrichment of

mafic minerals in the outer zones of the pluton and is supported by the presence of partially assimilated inclusions in the marginal rocks (Karner and Connors, 1971). The large, low areas of the pluton, some of which are indicated in Figure 2 as unsampled areas, may represent zones underlain by less resistant mafic inclusions or contaminated rocks.

Systematic zircon and zirconium variation gives further information on processes that operated in the pluton during crystallization (Karner and Helgeson, 1970). Average zirconium content decreases from 660 ppm in marginal granites to 170 ppm in the core. Zircon decreases in mean length from 0.24 mm in the margins to 0.13 mm in the core. Core zircons tend to be partially metamict and less elongate and have more complex crystal form than zircons of the margins. Frequency distributions of zircon characteristics show that marginal rocks contain zircon populations similar to the average population calculated for the pluton but with additional large crystals, and that core rocks have populations similar to the average but with large crystals removed. Broken zircon crystals and clusters of many crystals associated with magnetite, allanite-epidote, and sphene support the concept of concentration of early-formed zircon by mechanical movement. These comparisons and the enrichment of relatively fresh zircon in the marginal rocks and partially metamict zircon in the core of the pluton can be explained by a mechanism involving gravity settling. The concept of concentration by gravity settling is appealing because of the high specific gravity (4.6-4.7) of fresh zircon. Shaw's (1965) data show that large zircons 0.03-0.05 cm in length would settle at rates of 0.02 - 0.06 m/year. This rate seems inadequate since the granite magma would have to remain fluid in the margins, even the chill zone, for an increase of zirconium content of two or three times the average. Two additional factors may have been important in accelerating zircon concentration.

Zircon occurs in the outer rocks as clusters of many crystals associated with magnetite and sphene, both probably also of early crystallization. Clusters of zircon and magnetite 0.1-0.3 cm in diameter, as are common in the marginal zones, would have settling rates on the order of 0.3 m/year.

Convection may also have accelerated zircon concentration in the margins and depletion in the core by upward movement in the center of the magma chamber and downward flow along the walls. Shaw (1965) discusses such flow regimes in cylindrical magma chambers. The pluton's oriented mafic xenoliths which dip inward at angles of 30^{o}-40^{o} in the marginal rocks may parallel flow lines. Inward and upward flow has already been suggested to explain the enrichment of quartz in the core of the body and its depletion in the margins. Figure 4 schematically illustrates the magma chamber along the Tunk Lake section, assuming the above two aspects of flow as indicated by solid arrows. Zone boundaries, zirconium content, and areas of zircon enrichment and depletion are shown. The pattern is interpreted as a section through the lower part of two convection cells, as indicated by dashed arrows. Zircon accumulated where downward-moving magma was slowed and deflected inward. Shaw (1965) estimates average convective flow velocities for a somewhat smaller magma chamber to be about 10 m/year. Average velocities on the order of meters per year would have augmented downward movement of zircons or zircon clusters in the marginal areas and yet could have allowed zircons to settle where currents moved laterally.

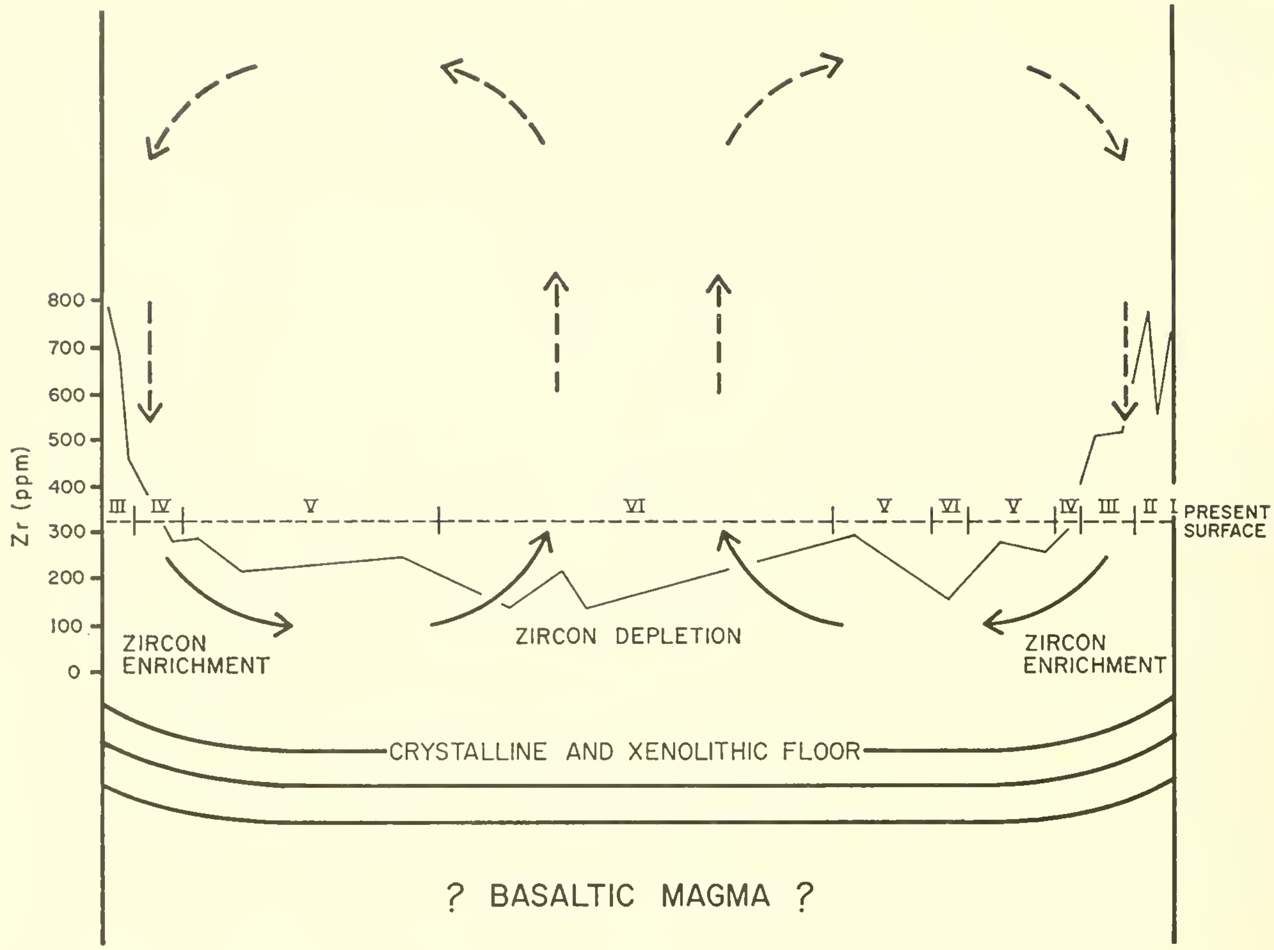

Figure 4. Diagrammatic illustration of convection cells in the Tunk Lake pluton postulated on the basis of zircon data (Karner and Helgesen, 1970). Zone boundaries and zirconium values from the Tunk Lake section (Fig. 2).

There are several possibilities for the pluton's shape and composition below the presently exposed level. If the convection cells of Figure 4 are minor, such as the secondary cells discussed by Shaw (1965), they may be underlain by other minor or major convection cells. If the flow patterns are the major convection cells of the Tunk Lake magma chamber, Shaw's (1965) treatment suggests the presence of a large underlying magma reservoir or a floor. The presence of a floor is assumed here because zircons would not be likely to accumulate at the base of a convection cell underlain either by a large magma reservoir or by other convection cells.

Melting-Anomaly Control

The pluton is within the Maine Coastal belt (Chapman, 1968b) and is related to several other circum-Atlantic groups of intrusions including the White Mountain plutonic-volcanic series of New Hampshire. Karner (1973, and in preparation) has suggested a New England melting-anomaly origin for the Maine Coastal and White Mountain bodies because of similar structural relationships and alkaline petrologic characteristics.

The Maine Coastal and White Mountain bodies (Fig. 5) are well known as two geographically separated groups of igneous complexes formed in the northern portion of the Appalachian tectonic belt. Twenty complexes in the Maine coastal region form a belt about 50 km wide which extends about 130 km northeast from Penobscot Bay and are assigned Devonian ages. About fifty separate centers of intrusive or extrusive igneous activity, composing about fifteen major central complexes and numerous minor and satellitic bodies, have been recognized in the Mesozoic White Mountain belt which is about 75 km wide and extends about 250 km from southern Maine and New Hampshire north-northwest to Vermont.

Similarities in structural relationships of both belts include: (1) crescentic, circular, and elliptical outcrop patterns which are generally interpreted as indicating ring dikes and stock-like intrusive forms; (2) occurrence of volcanic rocks and hypersolvus petrographic types suggesting shallow levels of intrusion and accompanying volcanic activity; (3) post-tectonic age in folded rocks of the New England Appalachians; and (4) reticulate patterns as noted by Chapman (1963, 1968a, 1968b) with plutons located at the intersections of inferred fracture sets. Chapman's (1968a) reticulate patterns are modified and related to major structural elements (Karner, in preparation). The patterns are NE and ENE for the Maine Coastal bodies. They are E-W and NNW for the White Mountain bodies in central New Hampshire, NE and ENE in southern Maine and New Hampshire, and E-W and NNE in northern New Hampshire (Fig. 5). In the Maine Coastal area the acute angle of the reticulate pattern of plutons is bisected by the trend of the proposed Maine Coastal Anticlinorium. The White Mountain belt extends NNW across the Rockingham Anticlinorium, Merrimack Synclinorium and the Bronson Hill-Boundary Mountain Anticlinorium. Most of the major plutons are in the Merrimack Synclinorium where the obtuse angle of the reticulate pattern is cut by the structural trend.

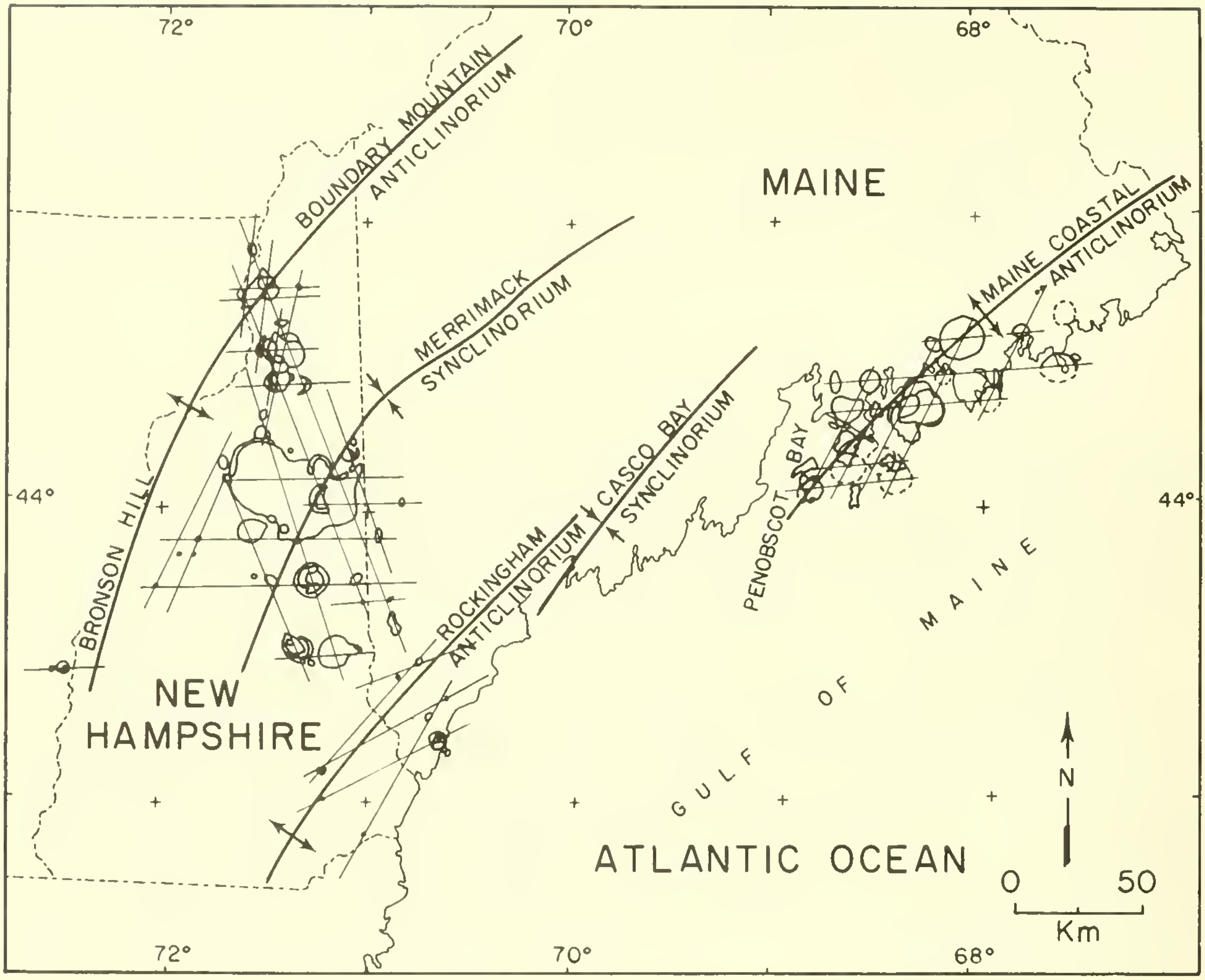

Figure 5. Igneous bodies of Maine Coastal belt in southeastern Maine and White Mountain plutonic-volcanic series in New Hampshire. Distribution of igneous bodies and rectilinear patterns modified from Chapman (1968a). Diagram from Karner (in preparation).

The reticulate patterns are believed to follow conjugate shear sets (Price, 1966) for anticlinal and synclinal structures.

In both belts a mildly alkaline rock series with the following major characteristics is present: (1) Early mafic phases including basaltic and gabbroic-dioritic rocks occur but are minor in abundance at exposed levels. (2) Initial felsic phases are quartz rich, occuring as Moat volcanics in the White Mountain area and as a distinctive chill zone in the Tunk Lake pluton. (3) Succeeding, felsic plutonic phases are predominantly intrusive and follow the sequence; undersaturated to saturated pyroxene-amphibole syenite, aegirine augite-hastingsite-riebeckite quartz syenite and quartz-poor granite ---- hornblende granite ---- hornblende-biotite granite ---- quartz-rich biotite granite and quartz monzonite. (4) The rock series are dominantly granitic with later subsolvus, mildly alkaline to calc-alkaline biotite granites and quartz monzonites by far the most common rock types. (5) Mineralogically, syenites and hypersolvus, quartz-poor granitic phases are most alkaline as demonstrated primarily by Na-rich minerals including Na-rich alkali feldspar, aegirine augite, hastingsite, riebeckite and astrophyllite. Distinctive minor minerals and accessories are fayalite, zircon, sphene, fluorite and allanite-epidote. (6) Chemically, the early felsic plutonic phases are relatively high in Na and K and low in Al and Ca compared to later calc-alkaline members and also contain trace element abundances characteristic of alkaline felsic rocks, such as high Zr.

In the Tunk Lake body the variations occur in a single zoned intrusion while in the White Mountain belt the rock types occur in composite bodies. Variation trends are parallel as shown on Figure 6, and with the similar petrographic characteristics discussed above, indicate a common genetic history. The Tunk Lake body is interpreted as a differentiated felsic cap produced from a convecting, mildly alkaline, granitic magma located above a mafic magma rising in a circular conduit. The White Mountain bodies and other Maine Coastal bodies are interpreted in a similar way by Chapman (1968b).

Morgan (1971, 1972) and many others (Karner, in preparation) have suggested a mantle-plume origin for the White Mountain series associating it with the New England Seamounts and present sites of Atlantic volcanism.

Karner (1973 and in preparation) makes the following major observations: (1) The Maine Coastal and White Mountain igneous bodies are structurally and petrographically similar, post-tectonic, mildly alkaline, plutonic-volcanic complexes in the northern portion of the Appalachian tectonic belt deformed during the Acadian Orogeny. (2) They have been emplaced at different times into markedly different geologic terranes as mafic bodies with granite caps. The Devonian Maine Coastal bodies are in a zone of gravity and magnetic highs in the proposed Maine Coastal Anticlinorium and the Mesozoic White Mountain bodies are largely in a zone of gravity and magnetic lows in the Merrimack Synclinorium. (3) In both areas, the bodies form reticulate patterns equivalent to conjugate shear directions of the enclosing major structures. (4) Both series have mildly alkaline petrographic characteristics and similar differentiation trends. (5) A mantle-plume (melting-anomaly) origin has been suggested by many workers for the White Mountain series relating the alkaline igneous activity to early stages in the separation of North America and Africa.

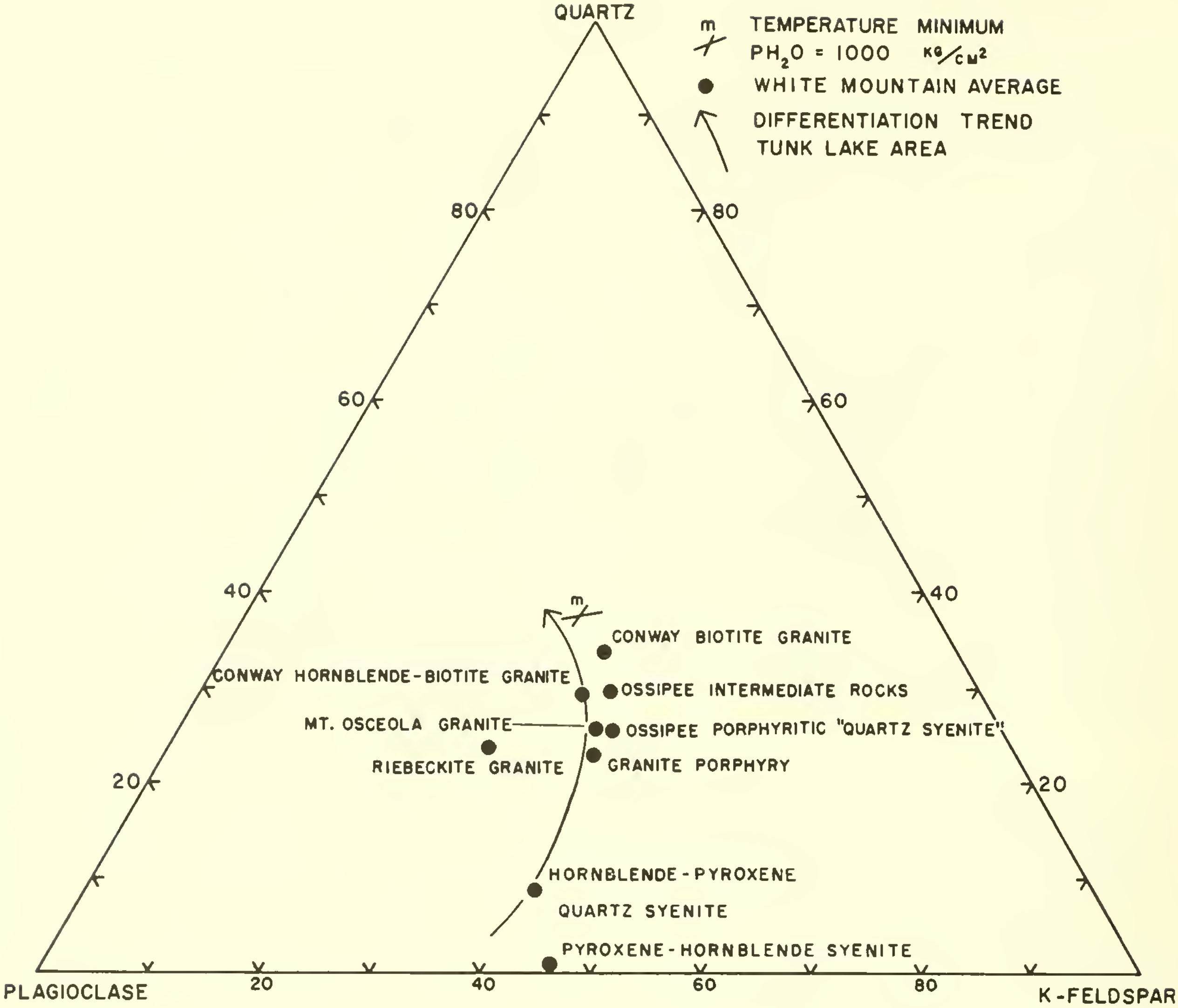

Figure 6. Variation of modal quartz, plagioclase and K-feldspar for the White Mountains plutonic volcanic series (Karner, 1968; Karner, and Bertram, 1972) and the differentiation trend of the Tunk Lake pluton (Karner, 1968) of the Maine Coastal series. The ternary minimum at 1000 kg/cm^2 is from work by Tuttle and Bowen (1958). Diagram from Karner (in preparation).

Karner (1973, in preparation) reaches the following tentative conclusions: (1) The Maine Coastal and White Mountain series have similar melting-anomaly origins. (2) Since the two series are emplaced into markedly different surface geologic terranes there is a deeper control of their formation, possibly the lower crust or upper mantle. (3) Since the patterns of the igneous bodies follow conjugate shear intersections and since they are inferred to have risen vertically from depth, the conjugate shears are characteristic of the rocks at depth. The shear intersections are present in a folded or undulating deep lithosphere as zones of weakness formed during the Devonian Acadian Orogeny and opened as magma conduits during Late Devonian-Carboniferous rifting in the Maine Coastal area and Mesozoic rifting in the White Mountain region. (4) A mantle source of alkali olivine basalt magma combined with partial melting of more felsic rocks is proposed to account for the petrography of the two series. (5) A New England melting anomaly related to deep folds or undulations or possibly other physical characteristics of the New England portion of the American lithospheric plate is proposed as an alternative to a melting anomaly related to present sites of Atlantic volcanism (thermal plumes).

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Itinerary

The field trip is designed so that we can study (1) the contacts with a variety of older rocks, (2) the unique chill zone, and (3) the gradation from hypersolvus, alkaline, pyroxene-granite to subsolvus, biotite, quartz monzonite in the concentrically zoned, Devonian, Tunk Lake pluton. Various features of the pluton including evidence for convective flow, typical rapakivi texture, and molybdenum mineralization will be examined. Please note that the field trip requires an 0.7 mi. hike from the north side of Catherine Mountain at elevation 470 ft. to the top at elevation 942 ft.

Mileage

0.0 — Assembly Point. Intersection of U.S. 1 and Maine 185 about one mile east of Sullivan, Maine.

Stop 1. Contact Zone of Southern Mixed Phase and Hornblende Granite. In this area, southwest of the Tunk Lake pluton, a mixed phase of diorite-gabbro cut by various types of granitic rocks is exposed as well as an elongate body of hornblende granite which may be related to the Tunk Lake body and is possibly a faulted segment of the southern margin of the pluton (Fig. 1). Note the following:

(1) The body of hornblende (aegirine augite?) granite east of the intersection. At small exposures 0.2 mi. east note the NNW orientation of compositional variants, elongate minerals, mafic xenoliths and shear? surfaces. Flow during emplacement and/or later tectonic deformation may have been involved in the formation of these features.

(2) The mixed phase exposed northwest of the intersection for about 0.3 mi. In the mixed phase, lighter rock types always cut darker, and darker types are always included in the lighter. Here, diorite-gabbro is cut by veins of hornblende granite as well as by a whitish, biotite- and plagioclase-bearing granitic phase. A hornblende granite dike? contains mafic xenoliths. Greenish-yellowish hornblende-aegirine augite granites and

syenites? (particularly to the northwest) are the most alkaline in appearance. Such rocks usually contain sodic-microperthite, aegirine augite, hastingsite, riebeckite and fayalite.

Return northwest on U.S. 1 toward Sullivan.

0.8 Approximate position of southeast contact of the North Sullivan pluton (Fig. 1). The uniform, medium-grained, biotite quartz monzonite of this pluton has been quarried extensively in the Sullivan-Franklin area. The contact zone is well-exposed on the west side of York Hill just to the southwest of the highway.

1.0 Turn north (right) on Maine 200. For the next 5.1 miles the road crosses the eastern part of the North Sullivan pluton. There are several small outcrops along the road.

2.9 Stop 2. North Sullivan Biotite Quartz Monzonite. Just east of the road a small outcrop of the typical rock is exposed. About 1.6 mi. northeast this rock is in contact with the marginal facies of the Tunk Lake pluton and is slightly altered suggesting that it is older. Note the following:
(1) Medium-grained biotite quartz monzonite of the North Sullivan pluton.
(2) Scattered small dark xenoliths.
(3) A narrow mafic dike.

Continue north on 200.

4.6 East Franklin.

6.1 Intersection of side road to northeast and approximate position of the north contact of the North Sullivan pluton with the Ellsworth Schist (Fig. 1).

6.9 Franklin. Turn northeast (right) on Maine 182.

7.8 Bayview Cemetery. Turn northwest (left) on road to Georges Pond.

8.1 Stop 3. Contact Zone of Ellsworth Schist and Coarse-grained Biotite Granite. A series of exposures of Ellsworth Schist about 0.1 mi. south of the road extends parallel to the road from the north side of the knoll behind the cemetery for about 0.3 mi. northwest to the covered contact of the coarse-grained, biotite granite (Fig. 1). The Ellsworth Schist is in a wedge-shaped area between at least three intrusions here (Fig. 1) and is strongly deformed. Note the increasing modification of the schist northwestward as shown by:
(1) Variation of foliation from well developed planar to contorted to weaker irregular foliation or almost massive structure.
(2) Progressive coarsening of texture.
(3) Increase of granitic vein material.

(4) Modification of amphibolite layers (probably originally basaltic sills) to contorted layers to aligned amphibolite boudins to unoriented amphibolite inclusions.

Return southwest toward Maine 182.

8.4 Intersection Maine 182. Turn northeast (left) on 182.

9.0 Approximate northern contact of Ellsworth Schist with the coarse-grained, biotite granite. For the next 4.3 miles the road crosses the southern part of this granite.

10.4 Stop 4. Coarse-grained Biotite Granite with Rapakivi Texture. Road cut on north side of highway just southwest of east end of old highway segment and opposite road to Duck Pond. The outcrop area of this rock type is cut by the younger Tunk Lake pluton about 0.8 mi. to the southeast (Fig. 1). Note the following:
(1) Typical coarse-grained biotite granite with perthite megacrysts up to 10 cm long.
(2) Occasional mantling of the perthite with plagioclase producing typical rapakivi texture.
(3) Mantling of perthite megacrysts in xenoliths.
(4) Scattered (and oriented?) mafic-rich areas in the granite.

Continue northeast on 182.

13.3 Approximate contact area of coarse-grained biotite granite to northwest, diorite gabbro to northeast and the Tunk Lake pluton to south (Fig. 1). For the next 9.2 mi. the road crosses the northern part of the pluton from margin to core and back out to the first stop at the eastern margin of the pluton. For succeeding stops, the field trip route will return on Maine 182 and end in the core of the pluton (Fig. 1).

13.6 Hornblende-aegirine augite of marginal zone II of the pluton is exposed on hills north and south of the road.

14.2 Approximate contact of zone II and zone III, hornblende granite.

15.0 Hornblende granite of zone III exposed along road at east end of Fox Pond.

15.3 Approximate contact of zone III and zone IV hornblende-biotite granite.

Catherine Mountain in the core of the pluton is southeast.

15.6 Approximate contact of zone IV and zone V biotite granite. The road is crossing the north flank of Catherine Mountain.

17.8 Tunk Lake. The road continues east across the core and back through the zones to the east margin of pluton.

22.1 Stop 5. Eastern Margin and Contact Zone of Tunk Lake Pluton. East end of road cut on Burke Hill at old highway cutoff access road. The biotite granite east of the pluton and the chill zone I of aegirine augite-magnetite granite gradational with zone II of hornblende-aegirine augite granite are exposed at this location (Fig. 1). Follow the blueberry-field access trail from about the middle of the old highway section eastward (parallel to the highway) about 0.3 mi. to near the eastern base of the N-S ridge marking the contact zone and resistant margin of the pluton. North and south of this location the contact zone is well exposed with a variety of mafic to felsic rocks present. Here and as you follow westward to the road cut note the following:

(1) The whitish tan medium-grained, hypidiomorphic granular, hypersolvus magnetite-aegirine augite granite of the chill zone I. The rock is strongly magnetic and this zone may account for the poor topographic rendering of this area on the Cherryfield 15' quadrangle map. Close to the contact some phases of the chill zone are unusual in that they are accessory mineral granites with no common ferromagnesian mineral, only about 1-2% magnetite, zircon, sphene and allanite-epidote in addition to feldspar and quartz.

(2) The older, medium-grained, biotite granite which is tan-lavender close to the contact and resembles the rock of the chill zone but, is clearly different in color and texture to the east farther from the contact.

(3) The transition of rock type I of the chill zone to type II, hornblende-aegirine augite granite. The grain size increases from medium-grained to coarse-grained. The color changes from whitish tan to the typical yellowish tan and green of alkaline granites. The color index increases from 1-3% to 6-8%, perthite increases from 50-55% to 70-75% and quartz decreases from 30-40% to 20-25%. The major changes occur in the first 30-50 m inward.

(4) The abundant, oriented, slab-like, mafic xenoliths which trend N-S to NW parallel to the contact and dip inward toward the center of the pluton at angles of 20°-45°. Detailed study suggests that xenolith abundance and orientation (strike) varies cyclically in zones parallel to the margin. The xenoliths are typically recrystallized mafic rocks consisting of about one-third aegirine augite and two-thirds albite-rich microperthite.

(5) Mineralization of N 80° W and other joint surfaces at road cut.

(6) Mafic dike exposed on north side of highway of road cut. The dike is several meters across, oriented E-W and almost vertical. It shows chilling at its contact with the granite and contains many xenoliths of granite in various states of partial assimilation. This and one or two other mafic dikes were the only ones found in the pluton.

Return west on Maine 182.

22.3 Approximate contact zone II and zone III, hornblende granite.

22.6 Stop 6. Hornblende Granite, Zone III. Outcrops of zone III north and northwest of two houses on the north side of 182 (Fig. 1). Note the following:
(1) Typical medium-coarse-grained hornblende granite of zone III. The rock has aegirine-augite only as tiny cores of hornblende crystals and slightly more quartz than type II. It is gradational with type II to the east.
(2) Abundant finer grained phase as dikes and irregular masses in the granite.

Continue west on Maine 182.

22.8 Approximate contact zone III and zone IV, hornblende biotite granite.

22.9 Stop 7. Hornblende-Biotite Granite, Zone IV. Outcrops north and east of cleared area north of the road (Fig. 1). Note the following:
(1) Typical medium-coarse grained hornblende-biotite granite of zone IV. It is gradational with type III to the east.
(2) Scattered buff and pink perthite grains mantled with white oligoclase.

Continue west on Maine 182.

23.0 Approximate contact zone IV and zone V,biotite granite.

23.1 Stop 8. Biotite Granite of Zone V with Rapakivi Texture. Road-cut on south side of road just west of Tunk cabins and just east of Washington-Hancock county boundary (Fig. 1). Note the following:
(1) Typical medium-coarse-grained, hornblende-biotite granite of zone V. The rock contains 30-35% quartz and 10-15% oligoclase. It is gradational with type IV to the east.
(2) Abundant pink to buff perthite grains mantled with white oligoclase to give typical rapakivi texture. A discontinuous rim of graphic quartz is present between rims and cores. The oligoclase is in optical continuity with exsolved albite in the perthite. Mantled feldspar typically occurs in this and adjacent zones and is believed to develop as the result of simultaneous hornblende to biotite reaction and feldspar exsolution.
(3) Development of fine-to medium- grained intergranular material (by recrystallization) producing hypidiomorphic granular to sub-porphyritic to seriate porphyritic to porphyritic textures in this zone.
(4) Several aplite dikes.
(5) Scattered, thoroughly recrystallized, medium-grained xenoliths 1-2 cm in diameter recognized as more or less equidimensional concentrations of ferromagnesian minerals. They are believed to be extensively recrystallized as a result of having been circulated several times through the magma chamber by convection currents in contrast to the less reacted xenoliths of the marginal rocks.

Continue west on Maine 182.

23.9 Approximate contact zone V and zone VI, biotite quartz monzonite. Type V is exposed on top of the small hill to the north and type VI on the hill's west side just ahead of us north of the road near the north end of Long Pond. Type VI is well-exposed on Round Mountain to the south.

26.4 Tunk Lake.

27.4 Stop 9. Biotite Quartz Monzonite Zone VI and Molybdenite Mineralization on Catherine Mountain. Starting at the northwest end of the cleared area follow an old trail 0.3 mi. northwest then 0.2 mi. southwest to the east side of Catherine Mountain and then walk 0.2 mi. west to the top (Fig. 1). Note the following:

(1) The typical, medium-grained to porphyritic biotite quartz monzonite of zone VI. It contains about one-third quartz K-feldspar and oligoclase with 3% biotite and magnetite. It is gradational compositionally and texturally with rocks of zone V.

(2) Abundant finer grained phases and aplite.

(3) The many molybdenite prospects around the top with mineralization in miarolitic cavities and in veins and on joint surfaces.

End.

BUCHAN-TYPE METAMORPHISM OF THE WATERVILLE PELITE, SOUTH-CENTRAL MAINE

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Introduction

This trip examines the recrystallization of the pelites of the Waterville Formation (Osberg, 1968; Heinonen, 1971) in response to a Buchan-type metamorphic event. The route parallels the regional strike of the Waterville Formation and participants may examine mineral associations from rocks of low chlorite zone through K-feldspar-sillimanite zone. Locations of stops are shown by numbers enclosed by circles in Figure 1. The area considered includes the following 15 minute U.S.G.S. Topographic maps: Pittsfield, Waterville, Vassalboro, Augusta, Gardiner, and Lewiston.

At low metamorphic grade the Waterville pelite consists of 1/4" to 1" interbeds of quartzite and quartz-mica slate. Some of the beds of quartzite contain a small amount of carbonate. In the low chlorite zone the pelite consists of quartz, plagioclase, muscovite, chlorite, $\pm$ hematite, and $\pm$ ferroan dolomite. The metamorphic grade increases southward, and isograds delineating the first appearance of biotite, garnet, andalusite-cordierite-staurolite, sillimanite, and K-feldspar-sillimanite have been mapped (Fig. 1).

Equilibrium and Phase Considerations

Equilibrium models for the Waterville pelite were discussed by Osberg (1971). K_D' values for Mg-Fe distribution between contacting cordierite and biotite are essentially identical in samples taken from compositionally different beds in a single outcrop. The proximity of the samples suggests that the external parameters (P,T, and possibly μ_j*) are essentially constant in the samples. The equality of K_D' in the samples indicates a close approach to chemical equilibrium.

Slight differences in composition of certain mineral species from grain cluster to grain cluster are interpreted as a manifestation of mosaic equilibrium, although it is not clear whether the mosaic equilibrium is a prograde process or whether it is due to retrogressive changes in external parameters or even to the effects of a later metamorphic event.

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* μ_j is the chemical potential of a mobile component.

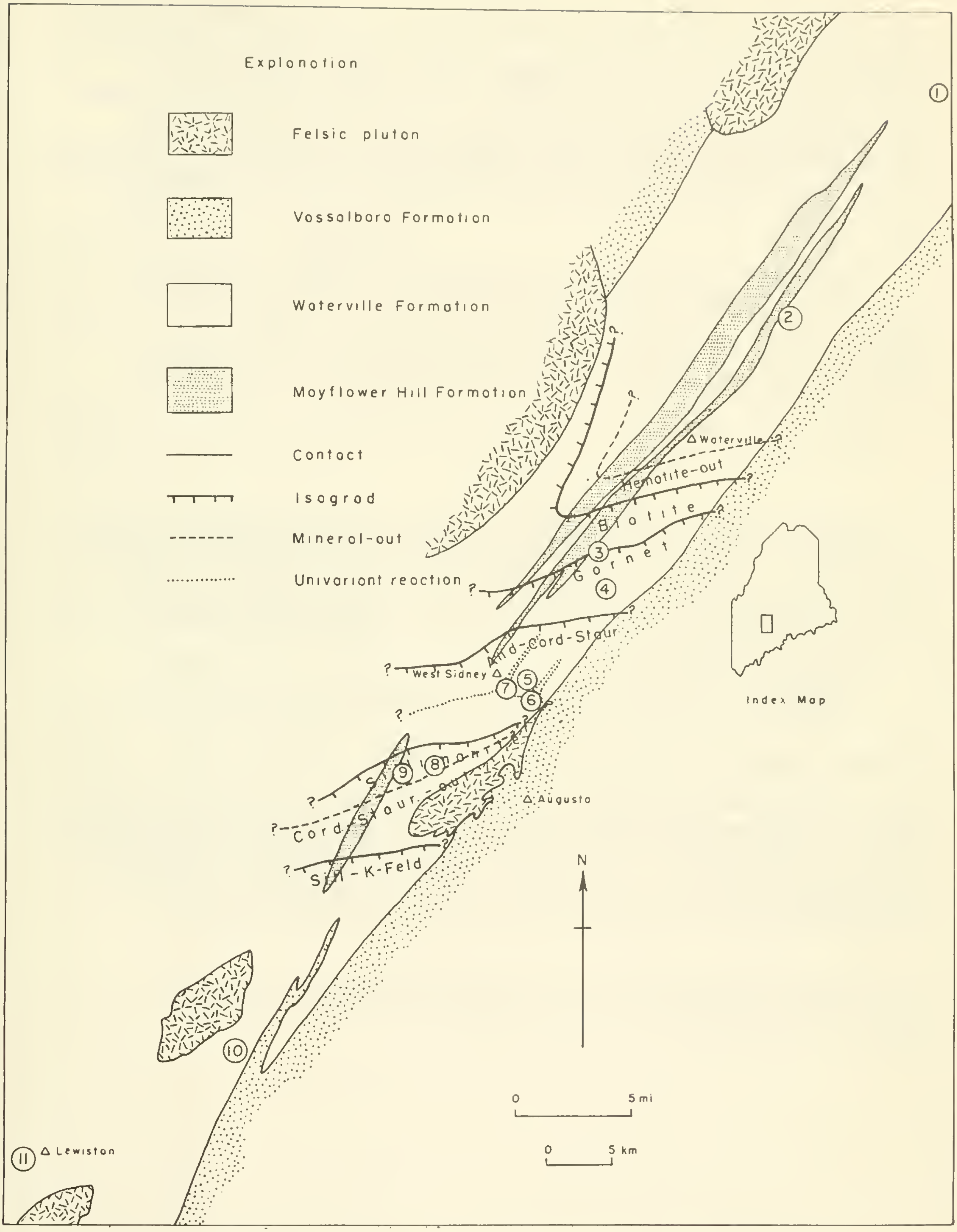

Figure 1. Metamorphic map of the Waterville Formation, central Maine.

The mineral relationships in an equilibrium assemblage should be consistent with the phase rule. At an arbitrary set of external conditions (P,T,μ_i), the number of phases is equal to the number of internally buffered components, $p = c_i$ (Korzhinskii, 1959; Thompson, 1959). In the metapelite of the Waterville Formation the maximum phase assemblage numbers twelve if an intergranular fluid that must have been present at the time of metamorphism is included. If eleven rock-forming metal oxides represent the maximum number of determining components, SiO_2, TiO_2, Al_2O_3, Fe_2O_3, FeO, MnO, MgO, CaO, Na_2O, K_2O, H_2O, SO_2, the maximum phase assemblage is (c_i+ 1). Implicitly the content of water and sulfur-dioxide is internally buffered.

The variance of the mineral associations at particular combinations of P and T is a function of the number of components and the number of phases. Because the number of components is large and in general the number of phases is few, most mineral assemblages and consequently most reactions are multivariant. Relatively few associations were encountered which could be considered to have a limited variance.

Reactions

Various isograds have been mapped in the field and these are indicated on Figure 1. The reactions responsible for the isogradic minerals are multivariate, and, therefore, the isogradic minerals crystallize not only in response to changing external parameters but also in response to variations in bulk composition. In such reactions, a new mineral is produced, and the preexisting minerals are not completely consumed, although they are modified in composition.

Conventionally, the biotite isograd is thought to be a result of the reaction:

chlorite + moscovite → biotite + aluminous chlorite + quartz + water.

Justification for this reaction in the Waterville pelite is based on x-ray data which indicate that in the chlorite zone the Al-content of chlorite is random, whereas in those rocks in which biotite is stabilized the Al in chlorite is systematic. At the biotite isograd the Al-content of chlorite is approximately 2.9 Al per 10 cations, and the Al-content of chlorite associated with biotite increases with grade as does the abundance of biotite.

The garnet isograd is drawn on the first appearance of almandine-spessartite. The composition of both biotite and chlorite change as garnet is produced by the following reaction:

biotite + aluminous chlorite + quartz → garnet + muscovite + slightly Fe-richer biotite + Mg-richer aluminous chlorite + water.

Apparently the variance of this reaction is such that the first appearance of garnet is strongly influenced by bulk chemical composition.

The andalusite-cordierite-staurolite isograd is drawn on the first appearance of one or more of the isogradic minerals produced by a reaction of the form:

muscovite + chlorite + garnet + quartz → biotite + water + one- or two-phase combinations $\left\{\begin{matrix}\text{staurolite}\\ \text{cordierite}\\ \text{andalusite}\end{matrix}\right\}$

Bulk composition strongly influences which of the isogradic minerals appear first. None of the reacting minerals is completely consumed during the reaction.

The sillimanite isograd delineates the first appearance of sillimanite and the disappearance of andalusite. Sillimanite, however is not produced by the direct conversion of andalusite. Instead sillimanite nucleates on biotite as mats of fibrolite, and the andalusite dissolves and is replaced by muscovite and quartz.

The sillimanite-K-feldspar zone is delineated by the association of K-feldspar and sillimanite. A reaction similar to that proposed by Guidotti (1963) is suggested for this region:

muscovite + sodic plagioclase + quartz → sillimanite + sodic microcline + less sodic plagioclase + water.

The dashed lines on Figure 1 represent the disappearance of hematite and the disappearance of staurolite and cordierite. Hematite initially reacts with ferroan dolomite to produce magnetite and Fe-poor dolomite:

hematite + ferroan dolomite → magnetite + Fe-poorer dolomite + CO_2.

In addition, hematite is reduced to magnetite. Because f_{O_2} differs from bed to bed, the hematite-magnetite buffer is reached at different temperatures in different beds.

The disappearance of staurolite and cordierite is interpreted to be due to the divariant reaction:

biotite + staurolite + cordierite + plagioclase = ilmenite + garnet + sillimanite + muscovite + quartz + H_2O.

Reactions of similar form, but involving staurolite alone or cordierite alone are also responsible for the disappearance of cordierite and staurolite and are more common then the c_i-phase reaction.

The association quartz + plagioclase + muscovite + biotite + chlorite + garnet + staurolite + cordierite + andalusite is interpreted to be univariant because phase considerations indicate it to be a (c_i+ 1)-assemblage. The (c_i+ 1)-phase assemblage is represented by the reaction:

ilmenite + muscovite + chlorite + garnet + andalusite + quartz = biotite + staurolite + cordierite + plagioclase + H_2O.

The (c_i+ 1)-phase assemblage is exposed in a single outcrop and the position of the reaction on Figure 1 is based on the compatibility of chlorite + garnet as opposed to that of staurolite + cordierite. The complex geometry of the dotted line representing the reaction is due to the fact that water, as a component, is internally controlled. It is not likely that P and T could vary sufficiently in a limited volume of rock to produce the configuration shown. Because the reaction is highly dependent on the number of moles of water, the variation in water content from bed to bed explains the map pattern of the reaction.

Conditions of Metamorphism

The temperature and fugacity of oxygen were estimated from the data of Lindsley (1963) for the (c_i+ 1)-phase association at West Sidney (Stop 7 in Fig. 1) using the compositions of magnetite and ilmenite. Extrapolation of Lindsley's data gives an approximate temperature of 500°C and an oxygen fugacity of 10^{-24} bars.

The maximum possible temperature of the hematite-out line (Fig. 1) is that of the hematite-magnetite buffer. Assuming that oxygen is an inert component and that its fugacity is of the order of 10^{-24}bars, the temperature of the hematite-out line is estimated to be 370°C using the relation of Eugster and Wones (1962).

The geologic relationships at the high grade end of the metamorphic spectrum (Stop 11 in Fig. 1) suggest that temperatures are somewhat below those needed to partially melt the rock. The temperature of the minimum melting curve for granite is estimated to be 685°C, suggesting that the temperatures realized were from 600°C to 650°C.

Total pressure is estimated to be in the range of 2.5 - 3.0 kb. The coexistence of cordierite and staurolite in middle grade associations indicates a range of pressures from 2.1 kb to 3.2 kb according to the experimental work of Hoschek (1969), Gangully (1972) and Siefert (1970). Extrapolation of the temperature from West Sideny (Stop 7 in Fig. 1) to the sillimanite isograd suggests a temperature of approximately 550°C at that isograd. Based on the work of Holdaway (1971), the total pressure is 2.8 kb.

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Itinerary

Mileage

0 Assembly point is in the parking space of the Carriage Inn and Restaurant, Somerset Avenue Exit of Route 95, Pittsfield. Starting time is 9:00 A.M. Breakfast is served in the Carriage Inn Restaurant from 6:30 A.M.

Stop 1. Leave cars in parking lot of Carriage Inn and Restaurant. Exposure is on the north bound entrance ramp to Route 95.

The outcrop contains three lithic types: black quartz-muscovite-chlorite-plagioclase-graphite-pyrite-pyrrhotite phyllite at the south end, grayish green quartz-muscovite-chlorite-plagioclase-magnetite-ferroan dolomite phyllite, and purple quartz-muscovite-chlorite-plagioclase-hematite-ferroan dolomite phyllite at the north end. The pyrrhotite in the black quartz-mica phyllite is monoclinic. The ferroan dolomite in the purple quartz-mica phyllite occurs as small porphyroblasts, as a major constituent of some quartzitic beds, and in quartz-ferroan dolomite veins.

The mineral associations are typical of the low chlorite zone. The presence of hematite-and magnetite-bearing beds in close proximity implies that f_{O_2} must have different values in different parts of the outcrop. In a gross way one might argue that the graphite-bearing assemblage is the most reduced, that the hematite-bearing assemblage is the most oxidized, and that the distribution of lithologies is a result of the mobility of oxygen, the graphite-bearing assemblage acting as an oxygen sink. The presence of interbedded purple and grayish green quartz-mica phyllite, however, argues for the differences in f_{O_2} as being related to sedimentational conditions.

You're not supposed to look at structural features on this trip, but in case you glance at them bedding and schistosity are essentially parallel and both are folded by asymmetric folds that are considered to belong to a late episode of deformation.

Return to cars. Turn left on Somerset Avenue and proceed onto southbound lane of Route 95.

11.9 Take Hinkley Road exit. Proceed across Hinkley Road and park

along southbound entrance ramp.

Stop 2. The outcrop is 500 feet south on Route 95.

Greenish gray quartz-muscovite-chlorite-plagioclase-magnetite-pyrite phyllite occurs in several outcrops. The outcrop is well banded with quartzose and pelitic beds alternating. A good number of the quartzose beds contain ferroan dolomite, and ferroan dolomite also occurs in quartz-carbonate veins. The veins are crudely zoned with quartz cores and carbonate rims. The mineral assemblage is that of the chlorite zone.

The schistosity is oriented at an angle to the bedding and is parallel to the axial surfaces of moderately open folds that predate those observed at Stop 1. The quartz-carbonate veins are folded and boudinaged.

Return to cars. Proceed south on Route 95.

22.8 Turn left onto Oakland Road (Rte 11 and 137) exit. Turn right onto Oakland Road.

23.8 Turn left onto Country Club Road and proceed south.

25.4 Stop sign. Turn right and proceed west.

26.0 Stop sign. Turn left onto Middle Road and proceed south.

29.3 Turn left onto Drummond Road and proceed east to overpass for 95.

30.3 Stop 3. Park along road. Outcrops are in field to the south.

Low exposure of greenish gray quartz-muscovite-biotite-chlorite-plagioclase-pyrite-ilmenite-magnetite schist. The biotite is porphyroblastic and its orientation is crossed with respect to the schistosity. Quartz, muscovite, and chlorite are very fine grained. This outcrop lies within the garnet zone, but garnet is extremely scarce. It occurs as small euhedral dodecahedra and has the following compositions: almandine = 67%, spessartite = 20%, pyrope = 7%, grossularite = 6%. Without the garnet, the rock looks much like those in the biotite zone.

Alternate beds of pelite and quartzite are well displayed. Schistosity is essentially parallel to bedding except in the noses of isoclinal folds. A second, later cleavage is also present.

Return to cars. Proceed eastward on Road.

31.2 Stop sign. Turn right on Sidney Road (Rte. 104) and proceed south.

33.2 Turn right on Lyons Road and proceed West.

33.8 Stop 4. Park along side of road.

Greenish gray quartz-plagioclase-muscovite-biotite-chlorite-garnet-ilmenite-magnetite-sulphide schist. Biotite and garnet are porphyroblastic, whereas other minerals are fine grained. Biotite has orientation that makes an angle to schistosity, and garnet occurs as small dark translucent dodecahedra. Several quartzose beds contain limesilicates: actinolite, plagioclase, and grossularite. A couple of beds containing quartz and carbonate are in the west end of the exposure. A prominent vein of quartz and calcite cuts the schist. These assemblages are typical of the garnet zone.

Bedding and schistosity are essentially parallel. Both are cut by a younger cleavage.

Return to cars. Proceed west on Lyons Road.

34.1 Cross overpass for Rte. 95 and turn left onto southbound lane.

37.9 Turn left into rest area. Lunch. Continue south on Rte. 95.

42.5 Turn right onto Belgrade Road (Rtes. 8-11-27).

44.1 Turn right.

44.3 Continue through intersection at North Augusta onto Bog Road.

47.4 At division of Bog and Middle Roads reverse direction and retrace route along Bog Road.

48.8 Stop 5. Park along side of road. Outcrop is on east side of road.

Exposure is a greenish gray quartz-plagioclase-muscovite-biotite-chlorite-garnet-staurolite-andalusite-ilmenite-magnetite-sulphide schist. The minerals are sufficiently coarse to be identified with a lens. Garnet, staurolite, and andalusite are porphyroblastic. Garnet occurs as small translucent red dodecahedra. Staurolite is honey brown and euhedral and in the outcrop has sharply etched, positive relief. Many crystals are twinned on {232}. Andalusite is white and crystals are glomeroblastic. In the outcrop the crystals standout as rounded "hob-nails". The ratios of MgO/(FeO+MnO+MgO in the ferromagnesian minerals are: biotite = 0.448; chlorite = 0.490; garnet rim = 0.079; and staurolite = 0.149. The mineral association is consistent with the andalusite-cordierite-staurolite zone and is interpreted to be a c_i-phase assemblage representing the (cd)-reaction.

Bedding and schistosity are essentially parallel. Garnet, andalusite, staurolite, and chlorite grow across the schistosity, without rotational features. A later slip cleavage cuts the schistosity.

Return to cars and proceed south on Bog Road.

50.7 Turn left on Old Belgrade Road at North Augusta.

51.1 Turn left on Middle Road.

52.6 Stop 6. Turn around in small parking lot and park facing south along side of road. Exposure is 600' north on woods road.

Outcrop consists of light gray quartz-plagioclase-muscovite-biotite-chlorite-garnet-cordierite-magnetite-ilmenite schist. Garnet occurs as small, euhedral, pink crystals. Cordierite occurs as large subhedral, ameboid crystals enclosing garnet, biotite, and chlorite. Cordierite weathers readily, so is recognized in the outcrop as depressions. The association is consistent with the andalusite-cordierite-staurolite zone and is thought to represent a (c_i-2)-phase assemblage. Because of the variance of the association, the compositions of the same mineral species from similar associations show a considerable and unsystematic variation with respect to metamorphic intensity.

Bedding and schistosity are parallel. Garnet, chlorite, and cordierite cut the schistosity.

Return to cars and proceed south on Middle Road.

54.1 Stop sign. Turn right onto Old Belgrade Road.

54.4 Stop sign. Proceed through North Augusta.

54.6 Stop sign. Turn right onto Belgrade Road (Rte. 11).

56.3 Stop 7. Park along east side of road.

Outcrop contains light gray quartz-plagioclase-muscovite-biotite-chlorite-garnet-staurolite-cordierite-andalusite-ilmenite-sulphide schist. Garnet is small and euhedral. Staurolite and andalusite standout in positive relief, whereas cordierite weathers out in negative relief. (A blind person could do a fairly adequate job by a sort of brail mineralogy.) This association is interpreted as a (c_i+1)-phase assemblage marking the shift in orientation of the tie-line garnet-chlorite to staurolite-cordierite. The compositions of the ferromagnesian minerals in terms of MgO/FeO+MnO+MO) are: biotite = 0.487; chlorite = 0.521; garnet rim = 0.090; staurolite = 0.151; and cordierite = 0.639. Two thin quartz-diopside-grossulairte-actinolite granulite beds are intercalated

with the pelite, and a large quartz vein cuts the outcrop.

Bedding and schistosity are parallel. Garnet, staurolite, cordierite, andalusite, and chlorite cut the schistosity and exhibit no rotational features. The bedding is folded by a large isoclinal fold. The schistosity is parallel to the axial surface of this fold. A late cleavage cuts the schistosity, and a late fold deforms bedding and schistosity. The late cleavage is parallel to the axial surface of the late fold.

Return to cars and proceed north on Rte. 11.

56.4 Reverse direction at Citgo station and proceed south on Route 11.

59.3 Turn right on Oakland Road.

60.6 Turn right onto Leighton Road.

62.2 Stop sign. Turn right onto Old Winthrop Road.

62.5 Stop sign. Turn right on Route 202 (Pine Tree Trail).

64.6 Turn right on Route 17 in village of Manchester.

65.4 Stop 8. Turn left into wood road and park. Caution of traffic.

Outcrop consists of light gray quartz-plagioclase-muscovite-biotite-garnet-sillimanite-magnetite-ilmenite-pyrrhotite schist. Garnet occurs as small euhedral pink crystals. Sillimanite occurs as small "maggot-like" mats of fibrolite. The assemblage is typical of the lower sillimanite zone. Cordierite or/and/ staurolite have been completely consumed by reaction with biotite. Several thin beds of quartz-anorthite-diopside-hornblende-garnet granulite are interbedded with the pelite, and within these beds the limesilicates have a zonal arrangement, indicating a limited mobility of components. Several small quartz veins cut the outcrop. On the south side of the road a vein containing prehnite in subhedral crystals cuts the schist (Please do not take specimens of prehnite.). Similar veins, although not common, are found througout New England, and may represent a regional zeolite facies metamorphic event (post Triassic?) that post-dates the higher grade regional metamorphism of New England.

Bedding and schistosity are essentially parallel and steeply dipping. A late cleavage cuts both bedding and cleavage.

Return to cars and proceed south on Route 17.

66.4 Stop sign. Turn right onto Route 202 (Pine Tree Trail).

68.2 Turn right onto "old" road through East Winthrop.

68.3 Turn right onto Case Road.

68.8 Stop 9. Turn around in driveway. Reverse direction and park on road.

Outcrop consists of light gray quartz-plagioclase-muscovite-biotite-garnet-sillimanite-staurolite-cordierite-ilmenite-magnetite schist. Sillimanite occurs as mats of fibrolite. Garnet occurs as small, red subhedral crystals. Staurolite occurs as small reticulated, yellow-brown grains set in large plates of muscovite, a relation similar to that described by Guidotti (1968). Cordierite can be seen only under a microscope. It occurs as small ameboid grains that include quartz and biotite. This association is regarded as a c_i-phase assemblage and represents the (Chl)-reaction. Several thin beds of quartz-plagioclase-hornblende-diopside-grossularite granulite are present. Again these beds show crude zoning due to the limited mobility of the chemical components. A pegmatite intrudes the pelite.

Bedding and schistosity are parallel and steeply dipping. Small asymmetrical folds belonging to the late episode of deformation deform both bedding and schistosity. A poorly expressed cleavage is parallel to the axial surfaces of the folds.

Return to cars and proceed south on Case Road.

69.2 Stop sign. Turn left.

69.3 Stop sign. Turn left onto Route 11 (Pine Tree Trail).

71.1 Turn right at Manchester onto Pond Road.

79.6 Stop sign. Turn right.

79.7 Bare left at fork at Litchfield Corner.

81.7 Stop sign. Turn right onto Routes 9 and 126.

89.1 Stop 10. Park along road. Outcrop is on east side of road. Caution of traffic.

Exposure consists of light gray quartz-microline-plagioclase-muscovite-biotite-sillimanite-garnet-magnetite-ilmenite schist. Sillimanite occurs as mats of fibrolite. Garnet occurs as small, subhedral grains. Microcline occurs as "chalky" white, irregular crystals. Chlorite is retrograde. Many veins of quartz and a few small pegmatites cut the schist.

Bedding and schistosity are parallel and steeply dipping.

Return to cars and proceed southwest on Routes 9 and 26.

92.6 Continue straight ahead on Route 126. Do not turn with Route 9.

97.1 Crossing light. Continue straight on.

97.5 Stop light. Continue on Route 126.

97.7 Stop light. Continue on Route 126.

98.1 Stop light. Continue on Route 126.

98.4 Stop light. Continue on Route 126.

98.6 Rotary. Turn left on Routes 202 and 11 through center of Lewiston and to Auburn.

99.0 Bridge over Androscoggin.

99.2 Stop light. Continue west through center of Auburn on Routes 202, 4 and 100.

99.4 Stop light. Continue west on Routes 202, 4 and 100.

99.5 Stop light. Turn left with Routes 202, 4 and 100 toward Maine Turnpike.

100.0 Large (not obvious) rotary. Follow signs for Routes 202, 4 and 100 toward Maine Turnpike.

101.2 Stop 11. Park along road. Caution of busy traffic.

Large outcrop containing light gray quartz-microline-plagioclase-biotite-sillimanite-garnet-muscovite-magnetite-ilmenite gneiss. Microcline is abundant and muscovite is relatively scarce. Garnet occurs as large subhedral crystals. Sillimanite occurs as abundant filrolite. Chlorite is retrograde. Two types of pegmatite are present: one has "fuzzy" boundaries with the gneiss, and the other has sharp boundaries. A thin basaltic dike, much younger than the metamorphism, also cuts the gneiss.

Layering and schistosity are parallel and gently dipping.

Return to cars. End of trip.

BEDROCK GEOLOGY OF NORTHERN PENOBSCOT BAY AREA

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Introduction

GEOLOGIC DESCRIPTION

An early Paleozoic Continental Collision

This field trip provides an opportunity to observe the evidence for an early Paleozoic continental collision between an ancient North America and an ancient Europe. The trip features a south-southeasterly traverse across major regional structures so that lithologic differences between rocks of equivalent ages can be emphasized and reasons sought for those differences. The traverse starts in Silurian and Devonian marine turbidite sediments of the Merrimack synclinorium in the ancient North American plate, crosses an anticlinorium of older rocks, and ends in a Silurian and Devonian synclinorium called the "coastal volcanic belt" made up of volcanic rocks and shallow-water sediments laid down on the ancient European plate. The differences in regional stratigraphy can be more easily appreciated if we can mentally remove the effects of polymetamorphism, many plutons, and a series of major faults. The lithologic differences are greater than those observed for facies changes and also include a change in the fragmentary fauna from those of North American assemblages on the northwest to those of European or Baltic assemblages on the southeast, and possibly different basement rocks with different metamorphic histories below the Paleozoic rocks. Our discussion will focus on the location and appearance of the suture between these ancient plates as evidenced by some major faults along the traverse.

Stratigraphy and Structural Geology

Basis for Age Assignments of Stratified Rocks

The geologic age assigned to each of the formations in figure 1 was deduced from meager paleontological evidence, from isotopic geochronology of whole rocks or separated minerals from the formations or from the plutonic rocks that intrude them, and from relative deformation or metamorphic styles of the formations. As the predominant contacts are faults, several ambiguities remain unresolved, and field mapping and isotopic studies continue.

Structural Blocks and Intervening Faults

Six regional structural blocks are recognized, and terranes and rocks representative of five of these will be seen on this trip. The sixth, or Islesboro block, is described in Stewart's trip A-6. Major faults bound each structural block. In geographic order, the Vassalboro-Waterville terrane of the northwesternmost block (Stops 1 and 2; fig. 2) is separated by the Norumbega fault from the terrane of Passagassawakeag Gneiss of Bickel (1971) and migmatite (Stops 3-5). The Long Lake fault separates lower grade terrane

Figure 1. TENTATIVE AGE ASSIGNMENTS OF ROCKS SEEN ON TRIP B-7

STRUCTURAL BLOCK	VASSALBORO-WATERVILLE BLOCK	PASSAGASSAWAKEAG BLOCK	BUCKSPORT BLOCK	PENOBSCOT BLOCK	ISLESBORO BLOCK	CASTINE - ELLSWORTH BLOCK
Devonian	Vassalboro 1/ Formation					Castine Volcanics 7/
Silurian	Waterville 1/ Formation; Mayflower Hill 1/ Formation					
Ordovician		"Stubbs" granite 2/ = Stricklen Ridge pluton	? — ? Appleton Formation of Cheney (1967) 3/; Bucksport Formation of Wing (1957); Schist of Rider Bluff ? — ?	Penobscot 4/ Formation		
Cambrian		Copeland Formation of Wing (1957) ?= Hogback Schist of Perkins and Smith (1925)			Islesboro 5/ Formation	? — ? Ellsworth Schist 8/
Precambrian		Passagassawakeag Gneiss of Bickel (1971)			Rocks of Seven Hundred Acre Island 6/	

Footnotes:
1/ Osberg (1968), J.R. Griffin, oral communication, 1973.
2/ R.E. Zartman, oral communication, 1974.
3/ D.G. Brookins, written communication, 1973.
4/ Boucot, Brookins, Forbes, and Guidotti, 1972; D.G. Brookins, written communication, 1971.
5/ D.G. Brookins, written communication, 1973.
6/ Brookins and Stewart, written communication, 1973.
7/ Brookins, Berdan, and Stewart, 1973.
8/ D.G. Brookins, written communication, 1973.

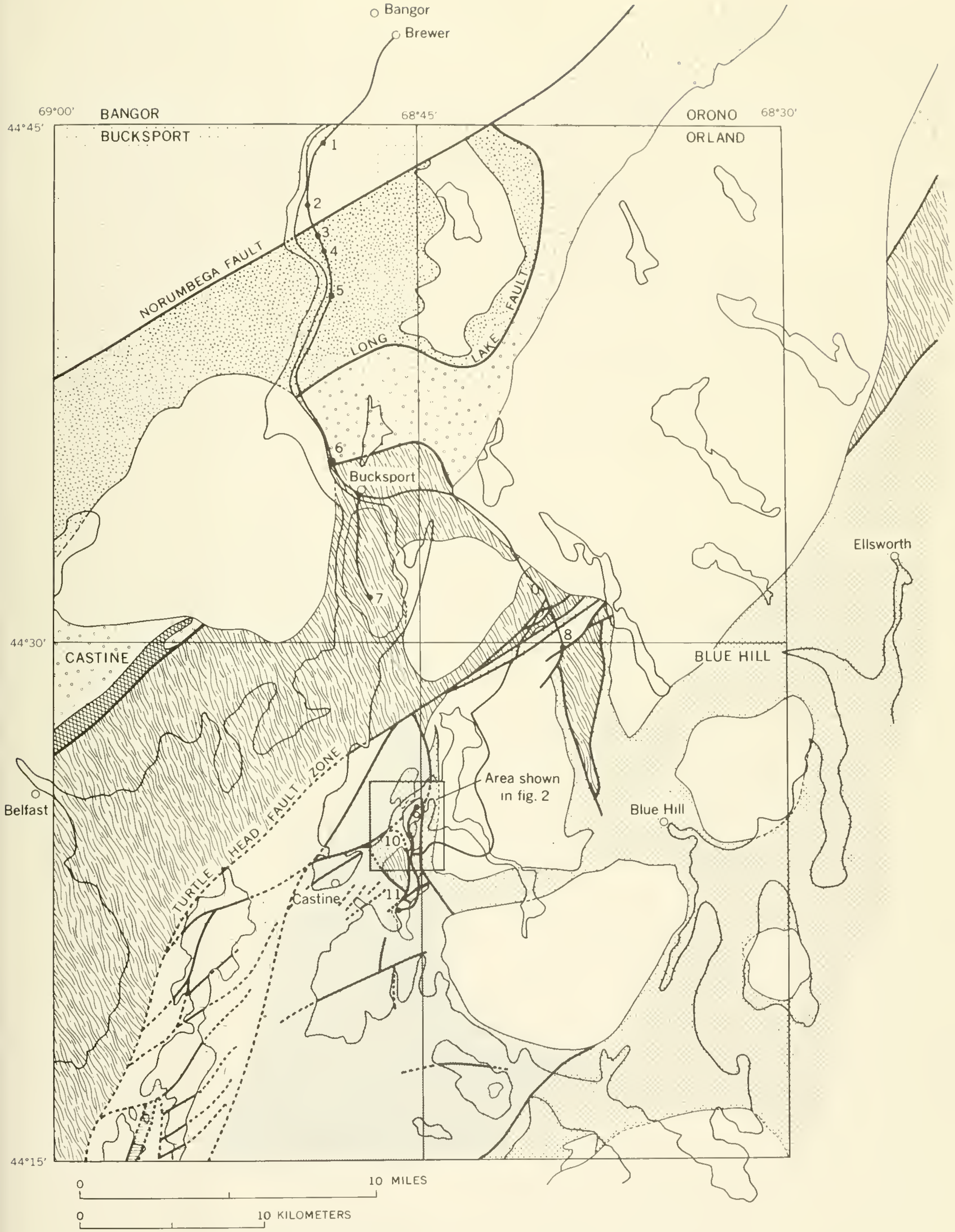
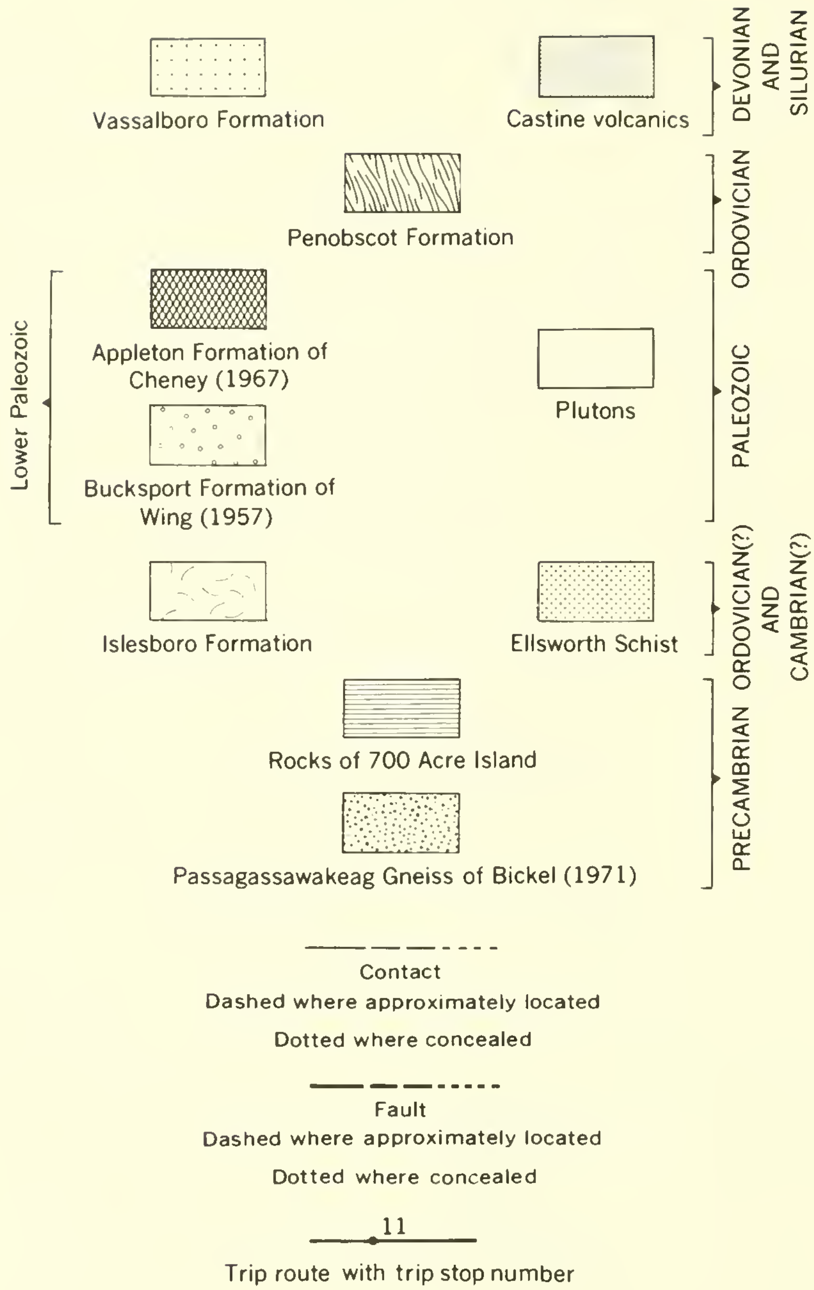


Figure 2. Geologic map of northern Penobscot Bay area.

EXPLANATION



of the Bucksport Formation of Wing (1957) from the Passagassawakeag. The Bucksport terrane is in turn probably separated from terrane underlain by the Penobscot Formation by a fault (Stop 6). The Penobscot block (Stop 7) is cut off on the southeast by a major strike-slip fault zone, the Turtle Head fault zone, which also removes all but a small sliver of the Islesboro Formation from along our traverse. The Castine Volcanics rest unconformably on Ellsworth Schist in the "coastal volcanic" block (Stops 9-11). The southeasterly limit of the "coastal volcanic" block is unknown. Several large faults in the Gulf of Maine involve Triassic and possibly Carboniferous rocks as well as lower Paleozoic rocks; all are unconformably overlain by Cretaceous and younger rocks of the Coastal Plain.

Not all of the faults are of the same age, and some have been active for long periods or were reactivated after earlier movements. On the trip, emphasis will be placed on evidence for the faults, about the nature of their movements, and about their effects on the rocks involved. Minor structures such as styles and sense of motion on minor folds, zones of shear and cataclasis, or cleaved, foliated, lineated, and polydeformed rocks all afford useful clues. Breccia zones, map patterns, and strong contrasts in metamorphic grade also provide evidence.

Regional Metamorphism

All rocks in the region are at chlorite grade of regional metamorphism or higher. A few Triassic dike rocks may be exceptions. Rocks of second sillimanite grade are found in the Passagassawakeag block, but, in general, the rocks outside the Passagassawakeag block and the zones of contact metamorphism associated with plutons do not exceed biotite grade. Within the contact aureoles, andalusite and cordierite are abundant, and sillimanite is relatively rare.

Several metamorphic episodes have been established within the region, but the exact number still is unknown. The youngest is not evident in the field in our map area. It is a Permian overprint of argon-bearing minerals reported by Zartman, Hurley, Krueger, and Giletti (1970), and its northeastern limit in New England probably reaches the southwestern part of our map area. Osberg (1968) attributed a Permian age to high-grade metamorphism in the Waterville area to the west, and probably the effects of this metamorphic event reach to the west shore of Penobscot Bay.

One of the attractive features of our mapped area is the opportunity to observe sedimentary rocks at low metamorphic grades and to unravel metamorphic events that have been obliterated by subsequent high-grade metamorphism. The northeastern limits of the high- and medium-grade Devonian Acadian regional metamorphism in New England must be west of Penobscot Bay. The high- and medium-grade metamorphism of the Penobscot Formation (Ordovician) south of Belfast and west of Islesboro in Penobscot Bay could be Acadian but possibly is older. The higher grade metamorphism of the Penobscot Formation is older than the Turtle Head fault zone, which places andalusite-grade Penobscot

Formation against chlorite-grade Islesboro Formation. There is a long history of Devonian movement on the Turtle Head fault zone; this movement began after the Castine Volcanics (Upper Silurian and Lower Devonian) were deposited, and stopped after the emplacement of the South Penobscot pluton (Lower Devonian) but before the emplacement of the Lucerne pluton (Middle Devonian). If Acadian metamorphism took place in Early to Middle Devonian time, Acadian metamorphic rock in the Penobscot Formation could be broken by the Turtle Head fault zone.

At least one other early Paleozoic metamorphic episode has been established in this area. Metamorphosed Ellsworth Schist (Cambrian? and Ordovician?) fragments occur on the unconformity below the Castine Volcanics (Stop 9). This unconformity is known to be pre-Middle Silurian at Ames Knob, North Haven Island, just south of the map area. No clasts of unquestioned Penobscot Formation have been observed on the unconformity, and no high-grade metamorphic rocks of any type have been noted, so the age of the metamorphism of the Ellsworth Schist is known only to be post-Cambrian(?) and pre-Middle Silurian. This metamorphic episode is not known in more complete lower Paleozoic sedimentary sections in New Brunswick, although a post-Middle Ordovician pre-Silurian unconformity is known in several places.

On the Islesboro block, the Precambrian foliated rocks of Seven Hundred Acre Island have been regionally metamorphosed to garnet-andalusite grade. They have been approximately dated at 750±150 m.y. by the Rb/Sr whole rock method (Brookins and Stewart, written communication, 1973). The rocks have been extensively retrograded and are cut by less foliated pegmatites that have been dated by two isotopic methods as also of Precambrian age (600±20 m.y. = Avalonian).

Within the Passagassawakeag block, three observed deformations are older than Late Ordovician, but it is difficult to assign a specific age to the high-grade metamorphism observed in these rocks. The age of the oldest rocks is Precambrian (D.G. Brookins, written communication, 1974). Gneiss in the Passagassawakeag block is more complexly deformed than the Precambrian observed in the rocks of Seven Hundred Acre Island and vicinity.

Plutonic Rocks

A variety of plutonic rocks of early Paleozoic age is discussed by Wones in the guide to trip A-7 and will not be reiterated here.

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Itinerary

Mileage	
0	Trip starts at Brewer, Maine, at east end of bridge across Penobscot River, Intersection of Route 15 and Route 1A. Turn south onto Route 15.
4.6	R.R. Crossing. Park cars parallel to R.R. track on south side of highway.
	Stop 1. Exposure of gray-to-green-weathering member of Vassalboro Formation, containing calcareous silts interbedded with rusty-weathering pelites. Pelitic units are more common to the south and east. This outcrop is typical of this formation in the Bucksport and Orono quadrangles. See trip A-3 for a complete discussion of these rocks. Compositional layering was probably sedimentary, but layer boundaries are now slip-cleavage surfaces striking N.55°E., and dipping northwest. Kink bands striking N.20°W. cause "waffling" of surfaces within the pelitic layer. Note the continuity of quartz veins. Rock is chlorite grade, containing white mica, chlorite, calcite, and quartz.
11.5	Outcrop on right (west) side of highway. Be careful of traffic.
	Stop 2. Exposure of Vassalboro Formation. Note that cleavage has further stripped out original bedding and compositional layers compared with those at Stop 1. The average pelitic unit is thinner. Note that quartz veins are stretched and broken. The dominant cleavage is N.55°E. and vertical. The metamorphic grade is chlorite; the rocks contain white mica, chlorite, quartz, and calcite. This outcrop is less than 1 km north of the Norumbega fault which is at the base of the small hills visible to the south. This fault strikes N.55°E., and strains related to that fault motion have affected the rocks at both Stops 1 and 2. (The name Norumbega is taken from the legendary Indian city which was the objective of 16th and 17th century explorers in Penobscot Bay (Morison, 1971).
12.2	R.R. Track. Trace of Norumbega fault crosses Route 15 here.
12.3	Outcrop on left (east) side of highway.
	Stop 3. Exposure of Passagassawakeag Gneiss of Bickel (1971). This material extends about 6-7 km (4 mi) to the northeast and at least 40 km (25 mi) to the southwest where it is exposed in the Belfast quadrangle. Most of this gneiss is a quartz-feldspar-biotite augen gneiss marked by feldspar porphyroblasts that are commonly arranged in thin lenses. The augen gneiss

is intruded by migmatite. A mafic-rich granodiorite (Winterport Granite of Trefethen (1941; 1944) intrudes the gneiss and migmatite and is strongly foliated by the pervasive N.60°E. foliation which is the dominant fabric. This in turn is intruded by another nonfoliated pegmatitic material. The grade of the block is uniformly at sillimanite grade, and evidence of second sillimanite grade is common. At this stop, vertical mylonite streaks strike N.55°E. These mylonites are vitric in appearance because of the exceedingly fine grain size of the crushed fragments.

The Norembega fault is 200 meters to the north and is a zone 300 to 400 meters wide. To the northeast, unmetamorphosed conglomerates with red matrices lie along the fault trace (Stoesser, 1966) and imply a post-Middle Devonian age for the latest motion on this fault. Displacements of metamorphic grade, the antiform to the northeast, and minor drag folds all imply left-lateral motion. Slicken-sides plunge N.55°E. with dips 25°-80° NE. The extension of this fault to the southwest is less certain, but it may connect with faults observed in the Casco Bay area (Hussey, 1968).

12.9 Budweiser sign on left (east). Turn right to Penobscot River boat landing.

13.1 Park cars at boat landing parking area and proceed to outcrop southwest of parking area.

Stop 4. Passagassawakeag Gneiss of Bickel (1971). Augen gneiss is intruded by the Winterport Granite of Trefethen (1941; 1944), which is foliate and intruded by pegmatite. Note that the scale of the rock units is in meters. This scale is observed throughout the area mapped as Passagassawakeag. The sequence of the units is the same as at Stop 3 and in all observed outcrops up to the time of writing, from oldest to youngest: (1) augen gneiss with schist, (2) foliate migmatite, (3) foliate granodiorite (Winterport Granite), (4) nonfoliate migmatite and pegmatite. Although dikes (sills?) of Winterport are most common in the northern section of the Passagassawakeag map area, they occur throughout the area.

13.3 Return to Route 15. Turn right (south).

14.1 Outcrop on left side of road.

Stop 5. Nonfoliate migmatite, "Stubbs" granite. This field is the site of largest mass of nonfoliate granite (migmatitic pegmatite) that has been found in this area. This material may be equivalent to the Stricklen Ridge pluton and part of

the same intrusive epoch. Zircons collected from this locality give a preliminary Pb^{206}/Pb^{207} age of 430 ± 10 m.y. (Zartman, oral communication, 1973). This implies that the Passagassawakeag Gneiss is older than Ordovician. Because of the multiple intrusions and deformations observed in that gneiss, it is presumed to be Precambrian. Wones believes that its composition and texture are more similar to gneiss in the Green Mountains than to the Precambrian rocks of Seven Hundred Acre Island (see trip A-6) or eastern Massachusetts. Further mapping combined with detailed isotopic studies is in process.

18.0 Sign "Welcome to Bucksport". You are now crossing the here-unexposed contact between the Passagassawakeag Gneiss of Bickel (1971) and the Bucksport Formation of Wing (1957).

This contact brings biotite-grade rocks adjacent to sillimanite-grade rocks, migmatite, and the Strioklen Ridge pluton, It truncates structures within the higher grade terrane, and a fragment of diorite is found along the contact. It also appears to truncate the contact between the lower grade Rider Bluff and Bucksport rocks. No basal conglomerates or other evidence was found for an unconformity, so a fault seems most likely. Geophysical data by Sweeney (1973) indicates a steep dip, and the map pattern is not compatible with a simple strike-slip fault. It is proposed that it is a normal or reverse fault bringing up the Passagassawakeag on the northwest against the Bucksport-Rider Bluff terrane on the southeast.

21.0 Turn left and park on right side of spur road. Return by foot to Route 15 and examine Bucksport Formation under powerlines. Walk along R.R. tracks to Penobscot outcrops, return to road, and walk along Route 15 to cars.

Stop 6. (1) Bucksport Formation of Wing (1957). The formation is characterized by interlaminated pelite and calcareous pelite, now quartz-biotite and plagioclase-amphibole bands. The compositional layers are related to the original bedding but may not be the original beds (compare with Stop 2). The compositional layers are a cleavage, and it has been folded, especially to the northeast. The overall lithology is grossly similar to that of the gray-to-green-weathering member of the Vassalboro Formation of Silurian and Devonian age. To the southwest, the pelitic members become much less common. Lithologic similarity need not imply chronological equivalence.

(2) Penobscot Formation. Rusty-weathering schist, sulfidic, and marked by bedding and composition lamellae on a centimeter

scale. No fossils have been found in it in northern Penobscot Bay, but similar-appearing rocks at St. John and St. Stephen, New Brunswick, contain Ordovician graptolites. The closer Benner Hill locality (see Osberg and Guidotti, trip A-3) is also Ordovician and may be equivalent to the Penobscot Formation.

(3) Waldo Granite of Trefethen (1941; 1944). The dikes seen crosscutting both the Bucksport and the Penobscot Formation are offshoots of the Mount Waldo pluton to the southwest. The contact of that pluton is beneath the Penobscot River and was drawn close to the eastern shoreline (Trefethen, 1944).

The contact between the Bucksport and Penobscot is disconformable to the southwest, so that it may be either a fault or an unconformity. Penetrative cleavage and regional metamorphic grade are higher within the Bucksport Formation than the Penobscot Formation, suggesting a fault contact.

The metamorphic grade here is high, because of the presence of the Mount Waldo pluton. To the northeast, the grade drops to biotite grade, but to the southwest the regional grade of the Bucksport remains high because of superposed metamorphisms

22.5 Turn right and cross bridge onto Route 1.

23.5 Continue straight (south) at Verona picnic area. (Route 1 bears to the right).

26.1 Turn left into Penobscot Rod and Gun Club. Club members may be shooting, so extreme caution is advised.

Stop 7. Penobscot Formation. (Please be certain to obtain permission from officers of the Rod and Gun Club before examining the outcrops). Note that the graded beds generally indicate tops of beds to the north in this highly contorted area: Sandy layers range from 1-10 mm, with occasional layers 100 mm thick. Pelitic layers are 4 to 400 mm thick. Andalusite, which has been badly retrograded to muscovite, is probably due to the contact aureole of either the Wallamatogus (to the east) or Mount Waldo (to the northwest) plutons, both of which crop out within 2 km of this locality.

26.2 Return to main road, turn right and return to Bucksport.

28.8 Verona picnic area. Continue straight (north) on Route 1.

29.8 Turn right on Route 1.

33.9 Turn right onto Route 15 towards Blue Hill.

39.0 Turn left at house 100 yards north of Amoco station. Walk 300 yards north through fields to small scarp, 3-4 m high.

Stop 8. Fault between South Penobscot pluton and Penobscot Formation strikes N.70°E. and dips 60° south. It is exposed here for 100 m. Many quartz veins are found within the Penobscot Formation along the fault trace. They contain terminated quartz crystals and have been the target of prospects such as the so-called "Annear Copper mine" which appears on the Orland quadrangle topographic map. The fault breaks the South Penobscot pluton. Near Castine and on northwestern Islesboro, Castine Volcanics are deformed by another fault within this fault zone, called the Turtle Head fault. Stewart postulates large strike-slip motion on the overall fault zone, but only normal fault motions are evident at this locality. The fault is intruded by the Lucerne pluton to the northeast, which indicates no movement since Middle Devonian time.

The South Penobscot pluton is highly variable in composition. Here it is a fine- to medium-grained foliate biotite tonalite with scattered remnants of Ellsworth Schist, mafic inclusions, and mafic intrusions. The age of the South Penobscot pluton is Early Devonian or younger on the basis of Rb-Sr and K-Ar dates and its contact metamorphism of the Castine Volcanics (Upper Silurian and Lower Devonian).

The Penobscot Formation is highly contorted and contains many quartz veins.

39.1 Return to Route 15 and turn right (north).

40.6 North Penobscot. Turn left onto Route 199.

41.6 White quartzite on right is a slice of Islesboro block caught in the fault zone.

45.7 Penobscot village, turn left onto Route 175.

52.4 Turn right on 175 and 176.

53.1 Bagaduce "River" with reversing tidal rip.

53.5 Turn right on Route 176.

56.1 Turn right on Ferry Road.

57.6 Farm yard belonging to Steve Woycke.

Stop 9. South side of Bagaduce Narrows. A lengthy woods walk will be necessary to reach the outcrops. This stop shows the unconformity between the Ellsworth Schist and the Castine Volcanics (fig. 3) and also shows that the Ellsworth Schist underwent a pre-Castine metamorphism (pre-Middle Silurain at Ames Knob, North Haven Island, south of Fig. 2). The unconformity initially had hundreds of feet of local relief

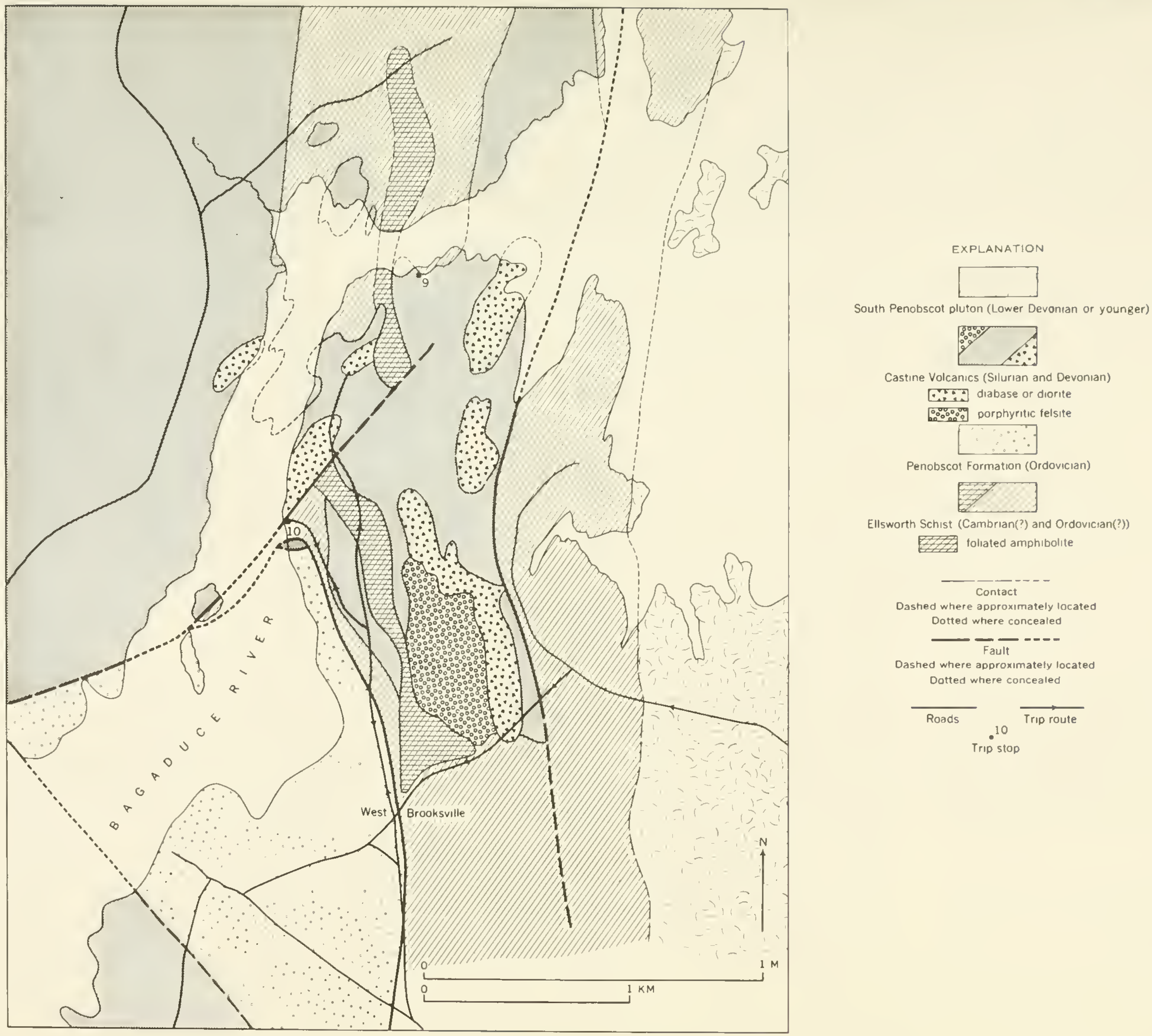

Figure 3. Geologic map in the vicinity of the Bagaduce River.

and has been folded subsequently. All the rocks were contact-metamorphosed to cordierite-andalusite hornfels by the South Penobscot pluton to the east (fig. 2), and metamorphic differences are evident between the Ellsworth and Castine rocks.

This unconformity was first recognized by Wingard (1958). The age of the Ellsworth Schist is not certain; it may possibly be Cambrian on the basis of Rb/Sr whole-rock isochrons by Brookins and Stewart. Typical low-grade Ellsworth Schist contains a quartz-plagioclase-muscovite-chlorite assemblage, and interlayered greenstones contain actinolite. Overall the Ellsworth Schist is notably feldspathic, plagioclase commonly being as abundant or more so than quartz. Bulk compositions are not as aluminous as is common in the Castine Volcanics. The leucogranite found as boulders above the unconformity is rare elsewhere in Ellsworth terrane, but could be a hypabyssal feeder equivalent to many crystal tuffs found within the Ellsworth. The contorted metamorphosed Ellsworth clasts above the unconformity show that extensive greenschist metamorphism had been imposed on the Ellsworth before its erosion. The metamorphic episode is known to be pre-Middle Silurian from the comparable unconformity 25 km southwest at Ames Knob, North Haven Island. A pre-Silurian post-Middle Ordovician orogeny has been identified elsewhere to the northeast in Maine or New Brunswick, but rocks as deformed and metamorphosed as the Ellsworth at this locality have not been noted. Relief on the unconformity is demonstrated by the great variety of immediately overlying basal Castine beds, from silts, sands, grits, to pebble and boulder conglomerates and by their rapid lateral variation. The trace of the unconformity, minor folds, and extensive diamond drilling indicate that the unconformity has folds with amplitudes of 5 to 75 m. The folds plunge gently south and are in the east limb of an anticline to the west.

Contact metamorphism by the South Penobscot pluton 1 km to the east has caused cordierite and andalusite knobs to form, particularly in the more pelitic Castine beds. Reaction of pyrite to pyrrhotite at high grade removed ferrous iron from hornblende in amphibolite beds and yielded anthophyllite near several sulfidic horizons that have been prospected.

Turn around and retrace route 0.9 mile.

58.5 Turn right.

59.0 Park at "I Wunda".

Stop 10. North of Lords Cove, West Brooksville.
A complex structure with strongly cleaved Ellsworth Schist in fault contact with Castine Volcanics and Penobscot Formation, although quite well exposed, is not completely understood because of inadequate knowledge of the third dimension.

The dirt access road roughly parallels the unconformity between Ellsworth Schist and Castine Volcanics. Basal Castine consists of arkosic grit with few quartz pebbles, and strikes approximately north, dipping gently (20°-30°) to the east. The attitude of the unconformity is unknown, but presumably it dips to the east. Typical contorted Ellsworth Schist is found beneath the unconformity. Near the end of the access road (by "Stewart" camp), and along the shore southward from where the road reaches the shore of the Bagaduce, the contorted Ellsworth is extensively sheared by a N.20°-30°E. steep cleavage. The cleavage is parallel to the trace of a major fault that drops diabase in the Castine Volcanics on the northwest side of the fault 50 to 100 m against Ellsworth.

Where exposed at the fishing shed at the end of the road, the fault strikes N.60°E. and dips 25°N., but slumping is suspected. Along the strike of the fault to the southwest on the southeast side of the northern Negro Island, the fault trace is strictly parallel with the very strongly developed fracture cleavage. Given the small vertical displacement of the fault, substantial strike-slip components are suspected to be necessary to cause the extensive cleavage observed. Although the fault breaks Castine Volcanics, it does not penetrate the South Penobscot pluton; its age is post-Early Devonian-pre-Middle Devonian.

Proceeding south along the shoreline, a representative, though highly sheared, section of Ellsworth Schist is exposed for several hundred m in front of two summer camps. A small N.70°W. fault dipping 45°-60°N. then separates sheared Ellsworth from 10-20 m of shattered Castine felsites. There are flutes in this fault that strike N.35°E. and plunge 40° NE. This fault traverses shears in the Ellsworth and is younger than the shearing episode. The Castine felsites are silicified and slightly pyritized and weather very whitish. They are distinct from Ellsworth rocks because, though foliated, they lack the lineations of the multiply deformed Ellsworth.

Passing over the plaque of Castine rocks, a N.80°W. fault dipping 45°-65° N. separates the Castine Volcanics from black pyritic phyllite of the Penobscot Formation. The fault plane is well exposed and also has flutes with as much as 2m of relief. These strike N.30°E. and plunge about 50°NE. The flutes are asymmetric and indicate upper plate movement to the east and north over the plastic phyllites. Contorted, highly sulfidic Penobscot Formation is exposed for hundreds of meters south along the shore.

The southernmost of the faults at this stop is exposed about 100 m horizontally and 10 m vertically, about equally above and below the high-tide line. In an easterly direction the fault

turns abruptly south and can be traced with some certainty 1.5 km through W. Brooksville and 1 km south. It is apparently dips quite steeply but is not exposed elsewhere. The presence of 2.5 km-long triangular block of Penobscot Formation surrounded by terrane composed of Castine Volcanics has been mind-boggling since its recognition, as the Castine has only been seen on Ellsworth Schist and never on Penobscot Formation. The fault motions observed at this stop are consistent with right-lateral fault motions. Two generations of faulting are clearly involved, and both have been contact metamorphosed by the South Penobscot pluton. A strike-slip fault zone, the Turtle Head fault, parallels the older fault at this locality, breaks the South Penobscot pluton and is cut by the Lucerne pluton (fig. 2). An explanation has been contrived that utilizes a long post-Castine history of north-easterly right-lateral strike-slip motion and the shouldering aside of blocks of Penobscot, Islesboro, and Ellsworth terranes by the diapiric intrusion of the South Penobscot pluton. It will be aired on the field trip if it is still viable and time and weather permit.

Turn around and retrace route.

59.4 Turn right.

60.0 West Brooksville, continue south on Route 176.

61.9 Turn right down steep paved road.

62.3 Park at Brooksville town landing.

Stop 11. Brooksville Town landing on Smith Cove.
The Castine Volcanics is exposed along the northeast shore of Smith Cove. A hornblende-rich diorite intruded into the section along faults has caused local contact-metamorphic effects. The diorite is sill-like locally and is 10 to a 100 m thick. Nonfoliated diorite like this is common the Castine Volcanics and is of late Castine age. Near Stop 9, similar diorite cutting Castine Volcanics was contact metamorphosed by the South Penobscot pluton.

The Castine Volcanics contains a variety of rocks. Igneous rocks include agglomerates, felsites, tuffs, and flows, varying from rhyolite to basaltic composition, most being siliceous. Sedimentary rocks include siltstone and mudstone rich in volcanic debris and some marine beds that contain calcareous or claystone concretions. One of the beds in the traverse along Smith Cove contains calcareous concretions metamorphosed to garnet grade by the diorite. Extensive

formation of cleavages has obliterated some primary sedimentary features, but there is little evidence for airfall deposition. Graded beds suggest that the section seen tops to the southeast. Beds 5 to 20 m thick are most typical and are highly variegated. All have attained at least chlorite grade of regional metamorphism. It has not been possible to establish a reliable stratigraphic column for the Castine Volcanics, and the impression gathered is that beds and even sequences of beds are of short lateral extent. This suggests several nearby sources and possibly some topographic influence. Sediments are at least as abundant as igneous rocks, though the igneous rocks dominate the present topography of the areas underlain by Castine Volcanics. The Castine Volcanics are latest Silurian and Early Devonian in age (Brookins, Berdan, and Stewart, 1973) and elsewhere contain fossils of Baltic aspect.

The section at Smith Cove has been broken by many small faults and shears that strike N.50°-70°E. and dip steeply. The strongest cleavage is N.30°-50°E. and steep, and there are at least three sets of kink bands: N.20°E., N.70°E., and N.70°W. Much of the tectonic disturbance of this section was cuased in Early to Middle Devonian time by the major strike-slip fault zone to the northwest and by the diapiric intrusion of the composite South Penobscot pluton to the northeast.

End of trip. Return to Route 1 via Route 176 to North Brooksville, 175 to Route 199 in Penobscot, 199 to Route 15 in North Penobscot, and Route 15 to Route 1.

Trip B-8 is the same as A-8. For the text and itinerary please refer to pages 126-135.

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